THE HYDROLOGICAL FUNCTION OF A MOUNTAIN VALLEY-BOTTOM PEATLAND UNDER DROUGHT CONDITIONS

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ABSTRACT

Mountain wetlands act as a sponge, storing water during wet periods and releasing water during dry periods. They are of particular interest as they have been shown to help mitigate downstream hydrological events, such as droughts and floods. Previous studies in northern wetlands have indicated the timing and magnitude of wetland runoff is inconsistent, with atmospheric and environmental conditions playing a key role in the production of wetland runoff. However, little work has been done to study the factors that influence flow between wetlands and streams systems in mountain valley-bottom regions. During the spring and summer of 2017, runoff dynamics of the Sibbald Research Wetland, a peatland in the Canadian Rocky Mountains, were analysed using a water balance approach and application of the Spence (2007) hydrological functions model. This model states that a peatland can store, transmit or contribute water to its outlet. An additional hydrological function, evapotranspiration, was added to this model to account for storage loss. Results show that the peatland was able to maintain outlet baseflow throughout the study period, despite a severe regional drought. Furthermore, the peatland transmitted water to its outlet when abundant ground frost was found in the upper 50 cm peat, whilst contributing water during the frost-free period. Additionally, large precipitation events initiated flows into peat storage which were quickly followed by runoff generation to the stream. Evapotranspiration occurred daily and accounted for the largest loss of storage from the system. This research indicates the importance of mountain peatlands in regulating streamflow during severe drought and during high precipitation events, as well at the importance of frozen ground and precipitation in determining the hydrological functioning of mountain peatlands. Moreover, this research underlines the need for further study in mountain peatlands across elevation gradients and for a variety of climatological and meteorological conditions as these controls on hydrological function may differ between peatland biophysical states and with atmospheric conditions.

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LIST OF ABBREVIATIONS

zw(t)	Water table depth
ΔS_{obs}	Change in observed storage
ΔS_{cum}	Change in cumulative observed storage
ΔS_s	Satursted storage
ΔS_u	Unsaturated storage
Δ	Slope of saturated vapor pressure
$\Delta\theta$	Change in volumtric soil water content
API	Antecedent Precipitation Index
API_d	Antecedent precipitation index for day d
B	von Karman contant
C_p	Specific heat capaciy for air
d	Displacement height of vegetation
E	Evaporation
e_a	Atmospheric water vapor pressure
e*	Saturation vapor pressure
E_A	Drying power
ET	Evapotranspiration
F_1:	Photosynthetically active radiation
F_2	Dependance of vapor pressure
F_3	Air temperature influence
F_4	Soil moisture influence
F_s	Sheltor factor
G_l	Lateral groundwater flow
G_l^{in}	Incoming lateral groundwater flow
G_l^{out}	Outgoing lateral groundwater flow
G_v^{net}	Net vertical groundwater exhange
h	Water table height
k	Decay factor
K	Hydraulic conductivity
K_c	Unit conversion coefficient
K_{in}	Incoming shortwave radiation
L_v	Latent heat of vaporization
LAI	Leaf area index
P	Precipitation
$P_{d:}$	Precipitation for day *d*
AET	Actual evapotranspiration
PM	Penman-Monteith
PT	Priestley-Taylor

Q^* .... Net surface radiation
Q_g .... Ground heat flux
Q_{in} .... Incoming streamflow
Q_{out} .... Outgoing streamflow
r_a .... Aerodynamic resistance
r_s .... Canopy resistance
r_s .... Stomatal resistance
S_y .... Specific yield
T_a .... Air temperature
u(z) .... Average wind speed at height *z*
u .... Wind speed at measured height
VWC .... Volumetric water content
z(t) .... Total soil depth
z .... Height of wind measurement
z_o .... Roughness height of momentum
z_{oh} .... Roughness height of water vaporγ: Psychometric constant
Δl .... Distance between monitoring wells
ρ_a .... Mean air density
ρ_w .... Density of water
ψ .... Soil suction at saturation

CHAPTER 1 - INTRODUCTION

1.1 Introduction

Mountain regions are often referred to as Earth's "water towers" (Viviroli et al., 2003), as they provide an abundance of freshwater to populations living within the reaches of their watersheds. Indeed, mountains are associated with large reservoirs of water, which may be present year-round. Highly variable weather systems, caused by extremes in mountain topography, result in deep snowpacks which commonly occur in alpine zones as precipitation generally increases and temperatures decrease with elevation (Millar et al., 2017). Glaciers may also characterize these zones. As temperatures increase throughout the spring and summer, melt from both reservoirs drains into valleys, on its way to densely populated lowlands. Mountain wetlands, fed by snow and glacial melt, are often situated in these valleys which are located at the geographic interface between mountain peaks and lowlands (Liu et al., 2004).

Among the common wetlands found in the mountain regions of North America, peatlands, particularly fens, are most abundant (Chimner et al., 2010; Morrison et al., 2015). The largest of these lies in valley-bottom positions (Morrison et al., 2015), and are able to hold large quantities of water due to the sponge-like characteristics of peat (Acreman and Holden, 2013; Bullock and Acreman, 2003). As a result, peatlands regulate water throughout the year, and display three main hydrological functions: they collect, store and discharge water (Black, 1997). Research on peatland hydrological function has been produced in a variety of northern regions including the Arctic, Canadian Prairies, Boreal Plain and Scandinavia (Roulet and Woo, 1988; Spence and Woo, 2003; Kværner and Kløve, 2008; Hayashi et al., 2016; Goodbrand et al., 2018), showing that these systems are capable of storing large amounts of water and releasing it at different times of the year.

Recent advancements in the study of peatland runoff regimes have also shown that hydrological functions occur as a response to thresholds dictated by peatland hydrological connectivity (Spence, 2010). While the characteristics and controls on peatland hydrological function have been well documented across a range of landscapes, little research has been done in mountain regions. Previous research has been limited to areas where variations in physiography and climate are small compared to mountain basins. Thus, it is unclear whether the characteristics and controls on peatland hydrological function known to other northern regions apply to mountain valley-bottom peatlands. Understanding the hydrological function of mountain peatlands located in valley bottoms is critical for improving storage-lag functions in current rainfall-runoff models (Bales et al., 2006).

1.2 Purpose and Objectives

The **purpose** of this thesis is to examine the hydrological role of large, valley-bottom, mountain peatlands in regulating streamflows. Through a variety of field measurements, the following **objectives** will be met: (1) characterize the storage-yield dynamics of a peatland; (2) explore the controls of these dynamics.

1.3 Literature review

The complexity of mountain wetland hydrological function is explained by their regional context, location within their catchment, as well as hydrological and internal mechanisms in the wetlands themselves. This literature review seeks to explain mountain wetland hydrological function by synthesising relevant literature in this area of study. In order to understand mountain wetland hydrological function, this review is divided into four sections including mountain systems, the hydrological function of mountain wetlands, the controls on wetland hydrological function, and finally the research gap.

1.3.1 Mountain systems

1.3.1.1 *Mountain physiography*

Mountain systems constitute about 27% to the earth's continental surface (Ives et al., 1997) and are immediately recognizable by their high elevations which significantly contrast the surrounding lowlands. A variety of different landscapes characterize mountain systems, which are often related to the elevation gradient. Distinguishing features include high elevation ridges, steep gradients, eroded rocky terrain and the presence of ice and snow (Barsch and Caine, 1984). The highest elevations are marked by the alpine zone, the area that extends above the tree line. The alpine has low temperatures, steep slopes and often strong winds, making it challenging for any form of life to inhabit these areas (Musselman et al., 2015). Sometimes these areas are also denoted by the presence of glaciers and long-lasting snow. During the warmest months, these frozen water reservoirs start to melt and water flows down slope, cutting into the bedrock to form channels (Hattanji et al., 2012). Evidence of past glacier erosion can be seen at these high elevations as well, including hanging valleys, cirques, arêtes and horns (Osborn et al., 2006).

Below the tree line, or the sub-alpine zone, lower altitudes feature an abundance of vegetation, although vegetation type is often limited by steep slopes and rocky terrain. In the subalpine, certain types of vegetation are more abundant than others. For instance, the sub-alpine zones of the Canadian Rocky Mountains are lined with coniferous trees such as Engelmann spruce (*Picea engelmannii)* and subalpine fir (*Abies lasiocarpa*; Osborn et al., 2006; Pomeroy et al., 2012). Immediately below the sub-alpine zone is the montane zone, which is marked by an abundance of vegetation and wildlife (McCutchan and Lewis, 2002). Here, climates are less extreme, and slopes are less steep. Depressions such as lakes, wetlands, tarns, as well as large amounts of glacial till found in U-shaped valleys create ideal locations for water from higher elevations to collect (Osborn and Bevis, 2001). Water from the alpine zones converge into larger streams and may form lakes and wetlands. The confluence of headwater streams is often characterised by large accumulations of poorly sorted sediment that has been transported by high stream flow, which can be seen during low flows (Wohl et al., 2017). The poorly sorted material of glacial till is known to store large amounts of water and serves as local aquifers in mountain valleys, feeding many streams, lakes and wetlands (Hood et al., 2006).

Finally, the base of mountain landscapes are characterized by the foothills zone (North and Henderson, 2010). Low and rolling relief marks the foothills. Here, vegetation is abundant and water reservoirs such as lakes and wetlands are plentiful (Morrison et al., 2015) due to the convergence of stored water from higher mountain areas (Wohl, 2017; Messerli et al., 2004; Viviroli et al., 2003).

1.3.1.2 *Mountain hydrological processes*

Mountains, often referred to as “water towers” (Viviroli et al., 2003), may represent as much as 95% of total regional streamflow and are the source areas for much of the world’s rivers (Liniger et al., 1998; Xu et al., 2009). For instance, four major European rivers, the Rhine, Rhone, Danube and Po have their headwaters in the Alps (Zierl and Bugmann, 2005). For the Danube and Rhine rivers, it is estimated that this mountain range contributes between 36%-95% of runoff respectively during the summer months (Viviroli and Weingartner, 2004; Diaz et al., 2003). In the semiarid regions of the western United States, it is estimated that mountain snowpack runoff accounts for 75% of annual streamflow (USGS, 2005; Bales et al., 2015). The range in runoff contribution varies throughout the year as the hydrological cycle is mainly driven by precipitation and melt, influencing vegetation composition and abundance, downstream water availability and biogeochemical fluxes at local and regional scales (Bales et al., 2006).

The extreme topography of mountain environments plays a key role in shaping mountain hydrological processes. The diverse climate of mountain regions is the principal factor in shaping the many diverse natural environments and determines the intensity of local and regional biological, physical and chemical processes (Barry, 1994). Elements that influence mountain climates include latitude, altitude and regional factors such as wind direction and ocean currents (Price, 1981).

The extreme changes in altitude that characterise mountain environments also shape climates as high and low peaks influence changes in atmospheric conditions. In general, temperature, air density, water vapour and carbon dioxide decrease incrementally with rises in elevation (Price, 1981). Furthermore, as humid air masses collide with these natural geographic barriers, condensation leads to precipitation in the form of rain at lower elevations and snow at higher ones. As colder air sinks deep into the valleys and rises once again (creating temperature inversions), humidity is further reduced. The effect that mountain topography has on climates produces some of the driest and wettest environments on earth (Cooper and Merritt, 2012). In general, however, the higher the elevation, the higher the likelihood that precipitation will fall (Millar et al., 2017). In high altitude areas where precipitation is abundant, rain and snow continually feed mountain reservoirs, not only in the snowpack but in, glaciers, lakes and wetlands and aquifers. In turn, these reservoirs actively contribute to headwater streams by releasing runoff in the form of overland and subsurface flow (McGlynn et al., 2004). As this runoff travels across the various mountain zones to lower latitudes, channels merge, creating higher order streams and recharging aquifers (Covino & McGlynn, 2007; Cooper et al., 2012). These aquifers are known to sustain hydrologically active areas such as lakes and wetlands (Devito et al., 1996; Grapes et al., 2006; Rosenberry & Hayashi, 2013).

1.3.1.3 *Mountain wetlands*

Wetlands are defined as areas saturated with water long enough to promote the development of water tolerant vegetation, hydric soils and water adapted biological activity (National Wetlands Working Group, 1988). They are located throughout the world's mountain ranges, from the temperate regions of the Alps and Himalayas (Koch et al., 2008), to the varied climates of the Andes and ranges in Africa (Islebe et al., 1996; Preston et al., 2003; Cooper et al., 2010). Although there has not yet been a complete inventory for the entirety of the Rocky Mountain range of North America, it is estimated that wetlands make up 2% of this region (Cooper et al., 2012). The ideal conditions for wetland development in mountains depend on both the physiography and water availability (Wilson et al., 2015). Precipitation, humidity and cool temperatures in these regions supply an abundance of water to support these landscape units. As these conditions are generally found at higher altitudes, this setting is where wetlands are commonly located (Cooper et al., 2012). In the San Juan mountains of Colorado, for instance, 90% of the 624 peatlands surveyed by Chimner et al., (2010) occurred at altitudes above 3000 m. However, wetlands were less abundant near the peaks due to steep slopes, creating unstable conditions for formation (Chimner et al., 2010). Indeed, ideal topography for these landscape units includes not only gentle slopes, but also the convergence of multiple streams found at mid and low latitudes (Westbrook et al., 2006). In Morrison et al.'s (2015) inventory, wetlands found in the front ranges of the southern Canadian Rocky Mountains were most plentiful at lower elevations in major U-shaped valleys where streams and lakes are also common. Additionally, many of the inventoried wetlands were highly concentrated in the lower foothills region. This is likely due to gently rolling topography, where valleys are abundant, and aquifers from glacial deposits are plentiful.

Wetlands are broadly classified into two groups: mineral wetlands and peatlands (Wilson et al., 2015). Mineral wetlands are found in areas where water collects at the surface, and where there is less than 40 cm of peat (National Wetlands Working Group, 1997). The water table may be found at or near the surface, but normally does not exceed more than 2 m above the ground. Soils are generally not well developed (Wilson et al., 2015). Mountain mineral wetlands include marshes, wet meadows and floodplains, and are common in intermountain basins across the United States and Canada (Cooper et al., 2012). Marshes occur at pond or lake margins, where they tend to have deep waters, and submerged aquatic plants grow tall. They may dry out periodically (Cooper et al., 2012). Wet meadows are groundwater fed, seasonally dry and are situated in sloping valleys. Floodplains are located in valley bottoms, where fluvial pathways coincide and are subject to flooding. In the Province of Alberta, mineral wetlands are most common in the grasslands and Prairie Pothole region (Gala et al., 2012). Peatlands are divided into two categories: bogs and fens. In order to be considered a peatland, these wetlands need to develop at least 40 cm of peat (The Canadian System of Soil Classification, 1998), a soil consisting of partially decomposed plant material including bryophytes, herbaceous vascular plants and woody debris (National Wetlands Working Group, 1998). The major difference between bogs and fens is that bogs are ombrogenous, whereas fens are minerogenous. Bogs receive water almost entirely from atmospheric sources and are thus hydrologically isolated from groundwater and surface water pathways. Bogs are more common in montane regions with a hypermaritime climate, like the coastal regions of British

Columbia and Alaska, where rain persists (Chimner et al., 2010). Primary production in these wetlands often exceeds decomposition, limiting the production of nutrients that many plants require, leading to the accumulation of peat. In contract, fens are minerogenous; they get their water from a variety of sources including groundwater, surface water, streams and precipitation. As a result, these fens are most common in continental mountain regions where they do not depend solely on direct precipitation for their water supply (Cooper and Andrus, 1994).

1.3.2 Wetland hydrological function

When Bullock and Acreman (2003) inventoried previous studies on the storage-yield dynamics of wetlands, they concluded that the same wetland could store and release water when different environmental conditions were met. Accordingly, studies on wetlands have observed various runoff regimes throughout the year (Spence, 2010; Phillips et al., 2011). In the arctic, subarctic and Precambrian Shield of Canada, the United States and Norway, valley-bottom fens and bogs have been found to regulate and generate streamflow based on internal and external factors governing the peatland itself and the surrounding environment (Roulet and Woo, 1986; Siegel, 1988; Branfireun and Roulet, 1998; Kværner and Kløve, 2008). For instance, Roulet and Woo (1986) observed that internal conditions such as frozen soils in spring and high evapotranspiration in summer respectively restricted and increased storage capacities of a low arctic peatland. As a result, the peatland was a poor regulator of flow in spring, generating large volumes of runoff, but a strong regulator in the summer. Kværner and Kløve (2008) found that variations in hydraulic conductivity of a peatland generated different volumes of groundwater runoff. Furthermore, Branfireun and Roulet (1998) noted that wetland runoff generation and attenuation was regulated by water table depth when the basin was dry and hydrological connectivity to the uplands when the basin was wet.

More recent studies have classified wetland storage-yield dynamics into three different hydrological functions: storing, contributing and transmitting water (Spence, 2007; Spence and Woo, 2003, 2006). According to Spence and Woo (2003) a wetland stores water when soil pore spaces fill up with water without exceeding capacity. The transmitting function occurs when a wetland relays runoff from an adjacent landscape unit to another (Spence and Woo, 2003). Finally, runoff contribution occurs when a wetland generates its own runoff (Spence and Woo, 2003). Empirical evidence has supported the idea of these three distinct hydrological functions. Spence and Woo (2003, 2006), Spence et al. (2010; 2011) and Goodbrand et al. (2018) describe storage, runoff transmission and contribution as a function of threshold. In their studies, wetlands with low water tables and low hydrological connectivity were able to store incoming water until water table thresholds were exceeded. When the threshold was met, excess surface or subsurface flow was generated and the wetlands started to transmit or contribute water to lower parts of the catchment. They would transmit water when runoff was equal or less than inflow. When internally generated runoff was greater than inflow, wetlands took on a contributing function. For instance, Spence and Woo (2006) noted that low water tables, caused by peak evapotranspiration in summer, allowed the wetland they studied to store rain and lateral inflow of water, without yielding runoff. The wetland began transmitting water when wet environmental conditions allowed for hydrological connectivity. Once inputs peaked, wetlands were able to start generating their own runoff and contribute to streamflow.

1.3.3 The controls of wetland hydrological function

Researchers attribute the state of wetland hydrological function to a series of thresholds dictated by internal and external factors (Spence, 2010). Internal factors include wetland soil and plant characteristics, as well as water table dynamics; whereas external factors include precipitation and upland runoff pathways. Hydrogeomorphic controls, which characterize both wetlands and their surroundings, also play an important role in wetland hydrological function. These factors influence the temporal shifts between functions as certain thresholds are met under hydrological conditions at different times of the year. The following section demonstrates the recent paradigm shift in threshold response and how these internal and external factors help shape wetland runoff regimes.

The focus of this section is to highlight the hydrological processes of mountain peatlands, more specifically fens, as they can be the most abundant type of wetland in these environments (Cooper and Andrus, 1994; Chimner et al., 2010). As there is a lack of literature on mountain wetland systems, this section focuses on knowledge gained from studying peatlands in the arctic, subarctic and prairie environments. However, as mountain peatlands are found at high elevations, with similar cold regions hydrology, these systems resemble those found in other geographies (Glenn and Woo, 1997; Zoltai and Pollet, 1983).

1.3.3.1 *Threshold responses*

For the last 40 years, much of the research done on catchment runoff has viewed runoff generation as a linear response between antecedent wetness and water inputs (Spence, 2010). However, recent advancements in catchment and hillslope hydrology now indicate that runoff generation acts as a non-linear, threshold mediated process (Spence, 2010). Although the principle of thresholds is not new, this paradigm has shifted from a function of continual storage accumulation or depletion, to a threshold dictated by soil heterogeneity, connectivity, hysteresis and non-steady state conditions (Spence, 2010). This new paradigm helps explain certain nuances in studies that did not conform to the old school of thought. For instance, Allen and Roulet (1994) noted that different catchment runoff mechanisms were generated depending on variations in soil type and bedrock geology. Shook and Pomeroy (2011) demonstrated the importance of catchment hysteresis when accounting for different storage and contributing areas within a basin. Furthermore, Zehe et al. (2005) demonstrated the importance of macroporosity (in addition to antecedent moisture) in determining thresholds. As a result, wetland thresholds are first and foremost dictated by internal factors, such as soil and plant properties (Waddington et al., 2015), water table dynamics (Waddington et al., 1993, 2015) and antecedent conditions (Shantz and Price, 2006; Waddington et al., 2015). These factors are then heavily influenced by external factors such as precipitation, surface and subsurface runoff. Finally, hydrogeomorphic controls are seen to heavily impact thresholds owing to both internal and external factors (Devito et al., 2005; Buttle, 2006).

1.3.3.2 *Internal controls*

1.3.3.2.1 *Soil characteristics*

Peat is the main soil type in peatlands. For peat to form and accumulate, there must be a constant supply of water, which may start in the form of a spring (Westbrook et al., 2006). As vegetation decomposes, peat accumulates, and peat density increases with depth (Ingram, 1983). The deeper the peat, the smaller the pore size and hydraulic conductivity (Whittington and Price, 2006). For example, in a study of summer runoff generation in a valley-bottom fen of southern Norway, where the peat reached a depth of 3.6 m, hydraulic conductivities determined via the Hvorslev method were found to decrease with depth, and the highest values were restricted to the upper 0.3m of the peat soils (Kværner and Kløve, 2008). The opposite can be said for peat pores in higher layers of a wetland. Less decomposed and compact peat exhibits larger pore size, which is responsible for transmitting higher amounts of water (Baird, 1997). Although previous models of peat density portrayed this soil matrix as diplotemic (Chanson & Siegel, 1986; Ivanov, 1981), where the soil layer consists of two different densities, newer research suggests that peat density, and therefore hydraulic conductivity, is much more complex (Hughes et al., 2000; Baird et al., 2008). For instance, in a raised bog of west Wales, lower hydraulic conductivities were found at the bog margins than towards its center (Baird et al., 2008). Furthermore, although hydraulic conductivities did decrease with depth, permeability of deep peat was high. Additionally, Shantz and Price (2006) showed that low hydraulic conductivity in a peatland could accelerate by preferential pathways attributed to the presence of woody debris, known as macropores or soil pipes. These sub-peat tunnels, often found at the interface of the peat base and the underlying substrate (Evans et al., 1999), have been found to facilitate large amounts of runoff. For instance, Jones (1979) found that 50% of stormflow runoff in a peatland was facilitated by soil pipes.

1.3.3.2.2 *Plant properties: evapotranspiration*

Many studies have indicated that evapotranspiration (ET), or the uptake of water by vegetation for photosynthesis, is the major cause of water loss in wetlands (Rosenberry et al.2004; Wright et. al, 2009; Spence et. al, 2011; Tardiff et. al, 2015). For instance, Peters et al. (2006) found that ET exceeded precipitation rates throughout the spring and summer, resulting in a lowering of the water table in between periods of rain; and Phillips et al. (2011) found that stream baseflow receded following high ET losses in the summer from a peatland in a Canadian shield basin. Indeed, high soil saturation over extended periods of time produce environments ideal for high-production micro-organisms and vegetation that might not been found elsewhere. As a result, evapotranspiration in wetlands is often greater than in other environments (Acreman and Mountford, 2009). For instance, reeds are capable of transpiring at least 14% more water than short grasses located in the prairies (Gilman et al., 1998). Additionally, variations in evapotranspiration rates vary with seasonal forcings. In a northern peatland of Ontario, for example, moss ET ranged from 200 W/m^2 during the growing season to less than 100 W/m^2 at the end of the summer as solar radiation gradually decreased (Admiral and Lafleur, 2007). Furthermore, it has been noted that ET rates seem to vary little between wetlands, as similar controls can be found across each system (Humphreys et al., 2006). Humphreys et al. (2006) examined six different wetlands across western and central Canada had ET rates that only varied by 0.13 mm/hr. The invariant ET was attributed to similar changes to leaf area index and mosses over time.

1.3.3.2.3 *Antecedent wetness conditions: water table dynamics and frost*

Fluctuations in water table depths vary annually and have a significant effect on storage, and ultimately runoff capacities at different times of the year (Whittington and Price, 2006). As peat hydraulic conductivity is low enough at depth to maintain year-round saturation, certain peatlands (more specifically fens) tend not to dry out completely during the year (Cooper et al., 1985; Evans et al., 1999; Chimner and Cooper, 2003). In general, local and regional water supplies are low and water tables depressed during late summer and autumn across arctic and subarctic climates (Roulet and Woo, 1986; Glenn and Woo, 1997). As these conditions do not drastically change over winter, antecedent wetness is low in spring, right before hydrologic activity increases (Roulet and Woo, 1986) As snowmelt begins and precipitation increases in spring, water tables are low enough to store excess moisture and attenuate flows (Roulet and Woo, 1986). Once thresholds are met, flow attenuations are minor until water tables decrease once again (Glenn and

Woo, 1997). Many studies indicate the importance of antecedent water table positions in the attenuation and generation of flows (Paavilainen and Päivänen, 1995; Spence and Woo 2003; Waddington et al., 2009). For example, Kvaerner and Kløve (2008) determined that antecedent storage in a Boreal fen was integral to decreasing peak flows. During a series of rainfall events, the highest runoff attenuations occurred when peatland water tables were at their lowest. Outflows were highest when storage was already nearing its threshold. In contrast, Bay (1969) determined that peak storm flow in a Minnesota peatland was three times larger when antecedent water table levels were near the surface than when they were 15 cm below.

Peatland storage thresholds are also determined by frost table depth. As frozen ground restricts the infiltration of water nearly completely, presence of frost restricts how much storage is available in spring when the snowpack starts to thaw (Hayashi et al., 2007). Hydraulic conductivity is also reduced during frozen periods (Guan et al., 2010). Roulet and Woo (1986) noted in the permafrost region of the Northwest Territories that spring snowmelt started before the frost table receded below the ground surface. As snow continued to melt, water levels in the fen would rise. The authors noted that spring was the only time of year where the peatland could not regulate flows, because the presence of frost limited storage capacities. Similar tendencies have also been observed in permafrost regions such as those in the Mackenzie River basin (Quninton and Hashi, 2005).

1.3.3.3 *External controls*

1.3.3.3.4 *Precipitation and runoff pathways*

Wetland inflows come from precipitation, surface and subsurface flows. Precipitation may fall directly on peatlands, contributing directly to storage, or may funnel into wetlands from the surrounding region via surface and subsurface flow (Owen, 1995; Millar et al., 2017; Spence 2010). Variations in latitude and seasonal temperature fluctuations dictate whether snow or rain will fall and create variations in storage (Price, 1981). Peatlands at higher elevations will receive more snow than those at lower elevations (Grünewald et al., 2014). Lower elevations will often receive more precipitation in the form of rain. For instance, the Front Ranges of the Canadian Rocky Mountains receive an annual average of 653 mm of precipitation, with 63% of this falling as rain.

Mountain hydrological processes are largely dominated by snowmelt or glacial melt with large amounts of initial surface runoff coming from the snowpack and glaciers (Osborn et al., 2006; Millar et al., 2017). There are two main mechanisms for which precipitation and melt can convert into overland flow: by infiltration excess (Hortonian flow) and saturation excess. Both of these mechanisms can occur along a same landscape element or hillslope, although are defined by the dynamics between storage thresholds and input quantity over time (Spence, 2010). Infiltration excess occurs as rainfall or melt exceeds the soil's infiltration capacity, and saturation excess is initiated when groundwater rises to the soil surface (McDonnell, 2013). Moreover, timing and magnitude of precipitation and meltwater discharge from the snowpack are the principle facilitators of saturation and Hortonian flow. McGlynn et al. (2004) showed how relatively small rainfall events with dry antecedent conditions resulted in a slow runoff response as water percolated quickly into the ground. However, a combination of higher water table levels and larger rainfall depth (amount) caused water table levels to rise above the ground surface where runoff was generated (McGlynn et al., 2004). This further suggests that the role of antecedent wetness conditions depends on rainfall characteristics (Castillo et al., 2003), where saturation-excess overland flow is more dependent on antecedent wetness conditions than infiltration-excess runoff.

Wetlands and other hydrological units also receive water from subsurface flows largely driven by groundwater recharge in the upper parts of a basin (Freeze and Cherry, 1979). Recharge rates depend on soil properties, the timing and magnitude of precipitation and antecedent wetness conditions (Goodbrand et al., 2018). For instance, in a study by Saxton et al. (1986), coarse-grained sandy soils were found to quickly recharge groundwater supplies due to low water holding capacities, whereas fine-textured soils retain larger amounts of water due to the high water holding capacity of small pore spaces. The timing and magnitude of precipitation and melt as well as antecedent conditions can also influence groundwater recharge (Redding, 2009). In a study on a catchment of the Western Boreal Forest, Canada, Smerdon et al. (2008) found that climate variations and water table depth largely affected groundwater recharge rates. In years of high snowmelt, groundwater stores were replenished, and higher water tables decreased the variability of groundwater recharge. Areas with thick unsaturated areas had high infiltration while water table depths were too low to facilitate water loss from evapotranspiration (Wittington and Price, 2006; Smerdon et al., 2008).

1.3.3.4 *Hydrogeomorphic controls*

Another important factor involved in wetland hydrological function is landscape physiography, referred to as hydrogeomorphic controls, which describes regional hydraulic connectivity. Various conceptual models of landscape position have attempted to describe the significance of landscape physiography on wetland hydrological function. For instance, Devito et al. (2005) uses a hierarchical sequence to understand the importance of physiography on groundwater flow patterns: climate- bedrock geology- surficial geology- soil type and depth. Similarly, Buttle (2006) suggests a three-component approach to understanding runoff production and flow through a catchment. In his T^3 template, Buttle (2006) proposes that landscape physiography is composed of topography typology, and topology. These three components characterize hydraulic connectivity, the capacity to transfer water from one part of the landscape to another (Bracken and Croke, 2007). Landscape topography demonstrates the spatial extent and boundary of individual landscape units and their contributing areas, and highlights their physical characteristics, including shape and size (Buttle, 2006; Phillips et al., 2011). Catchment typology, on the other hand, is the ability of different landscape elements such as soil, vegetation, geology and slope to generate runoff (Buttle, 2006; Spence and Woo, 2006). Finally, topology outlines the placement of these physiographic units within the landscape and their contributions to one another (Buttle, 2006; Spence and Woo, 2006). These three physiographic components help explain landscape threshold response. For example, at the wetland scale, a wetland's storage threshold is influenced by the topography, typology and topology of a unit and region (Spence, 2010). In another study, Guan et al. (2010) showed upland wetlands had minimal storage capacities along their edges compared to the ones in valley-bottom positions. Wetland margins became areas of runoff generation where as the centers of headwater wetlands became sinks for water, which was largely attributed to the shape and size of the wetland and its surroundings. Similarly, when evaluating the water budget of a terminal wetland, Spence et al. (2011) found that the wetland collected runoff in sequences, largely attributed to the relative physiology of the surrounding landscape units coupled with timing of rainfall, snowmelt and runoff generation rates in the headwaters of the catchment.

The placement of a wetland within a catchment is correlated with the amount of discharge it produces (Spence et al., 2011), as streams and ultimately wetlands receive water from all areas of the larger catchment (Goodbrand et al., 2018). However, it has been suggested that wetland inputs should be considered, not by adding direct contributions from each individual landscape unit, but by recognizing that landscape physiography plays an important role in regulating how much water is distributed to each basin (Spence and Woo, 2006). For example, in a study performed by Spence and Woo (2006), uplands and valleys generated very different amounts runoff during hydrological events of different magnitudes. During small, intensive events, runoff contributions were restricted to upland bedrock environments, limiting contributions in the valley bottom aquifers. However, during longer, less intensive events, bedrock uplands were found to generate water, whereas the valley bottoms tended to store and contribute water to the surrounding landscape.

1.3.4 Research gap

It has been hypothesized that wetlands have three important hydrological functions: they can store, transmit and contribute water (Spence and Woo, 2003, 2006; Spence et al., 2011). Much of the research on wetland hydrological function has taken place in the Arctic and subarctic Canadian Shield (e.g. Waddington et al., 1993; Spence & Woo 2002, 2003, 2006), Boreal Plain (e.g. Goodbrand et al., 2018) and Scandinavia (e.g. Kløve and Bengtsson, 1999; Kværner and

Kløve, 2008), some of which are peatlands. While these studies provide examples of wetland hydrological function, they were conducted in areas of much flatter relief, and different climatic and geological settings than is found in mountain environments. Therefore, questions remain on whether the understanding of northern peatland hydrological function is transferable to mountain peatlands. Specifically, it is relatively unknown how groundwater and surface storage, runoff sequence and water flow celerity from headwaters to valley-bottoms affects mountain wetland storage and ultimately runoff generation. Such research is critical in understanding how and when mountain reservoirs store and release water downstream, especially under low-flow or high-flow conditions. With increasing need for regional and national water security, this research will contribute to quantifying water supplies important for municipal, Indigenous, agricultural and natural zones that rely on Canada's water towers.

1.4 Thesis layout

This thesis has been written in a traditional, chapter style, format. The following chapters outline the methods used in the research, the research results, interpretation of the results and concluding remarks.

CHAPTER 2 - METHODS

2.1 Study Site

This research was conducted at the Sibbald Research Wetland, located in the Front Ranges of the Canadian Rocky Mountains, about 70 km West of Calgary (Figure 2.1). This ~0.71 km^2 valley-bottom peatland is surrounded by forested foothills reaching 1650 m above sea level, where marine clays and an unknown amount of alluvium underlie spatially variable thicknesses of peat. Hummocks and hollows characterise the land surface, while five streams originating in the uplands flow into the peatland. These include the southeast, east, northeast, north and west streams. The area of the watershed is 9.3 km^2. Additionally, Bateman Creek, a third order tributary of Jumpingpound Creek, drains the peatland, eventually flowing into the Bow River. Due to intensive flooding in the spring of 2013, beaver dams that had created large ponds along on the northern end of Bateman Creek were breached. The beaver population had yet to recover to its pre-2013 flood level in the peatland at the time this research was carried out. Thus, beaver activity was minimal during this study, compared what is normal at this site (Westbrook and Bedard-Haughn, 2016). Beaver activity was limited to the south and southeast parts of the peatland during the study.

The climate at the Sibbald Research Wetland is characterised by warm, dry summers, and cold, snowy winters. From 1981 to 2010, air temperatures recorded at the University of Calgary Biogeoscience Institute averaged -6.1ºC in January, and 14.5ºC in July. Average precipitation during this period was 639.3 mm, with 40.1% falling as snow. From June though August, total average rainfall was 255.1 mm. The area is also characterized by frequent freeze-thaw periods in winter due to chinooks that come from the west.

The peat that characterises this valley-bottom has variable thicknesses. In general, peat thickness is shallowest (<50 cm) at the northern edge of the peatland, and deepest peat (1-5 m) towards its center. Furthermore, the upper 50 cm of peat is mainly composed of sedges, whereas the deeper layers in the northern part of the peatland (50-130 cm below the peat surface) are mainly dominated by *Sphagnum* spp. mosses (Wang, et al., 2016). Additionally, thick stretches of willow (*Salix* spp.) dominate the northern end of the peatland, with slightly lower density than in the southern end.

As Sibbald Research Wetland has been an active research site since 2006 (Westbrook and Bedard-Haughn, 2016), several long-term installations have been put in place around the site to gather data (Figure 2.1). A large well network, consisting of 51 wells along 11 transects, covers about three quarters of the peatland, with transect 1 located in the north end. In addition to being used for monitoring water table elevations (Karran et al., 2018), this well network serves as a grid for measuring peat and frost table depths. The peatland is also equipped with a meteorological station, which is located near the peatland outlet. It collects hydrometeorological information year-round as part of the Canadian Rockies Hydrological Observatory (http://www.usask.ca/hydrology/CRHOStns.php). Measurements obtained from this station supports water budget calculations. These measurements include: air temperature (°C) and humidity (%) with a Rotronic HC2-S3 temperature & humidity probe, wind speed (m/s) with an RM Young 05103-10 wind monitor, soil temperature at 25 and 50 cm depths (°C) with a Campbell Scientific 107B temperature probe, rainfall (mm) with a Texas Electronics TE525 tipping bucket rain gauge, net radiation (W/m^2) with a Kipp and Zonen NR Lite net radiometer, snow depth (m) with a Campbell Scientific Canada SR50-45 Sonic Ranger, soil moisture (%) with Campbell Scientific CS616 soil moisture probe at 25 and 50 cm depths, soil heat flux (W/m^2) with a

Radiation Energy Balance Systems HFT3 soil heat flux plate, soil temperature (°C) with an Omega Type E thermocouple, incoming shortwave radiation (W/m^2) with an Apogee SP110 pyranometer, and atmospheric pressure with a Solinst Barometer Edge. Signals are recorded with a Campbell Scientific datalogger. This station operates at 5 s intervals and data are amalgamated at 15 min intervals (Westbrook & Bedard-Haughn, 2016). Finally, stilling wells placed along the five inlet streams described above and Bateman outlet, record stream stage when equipped with Solinst level loggers.

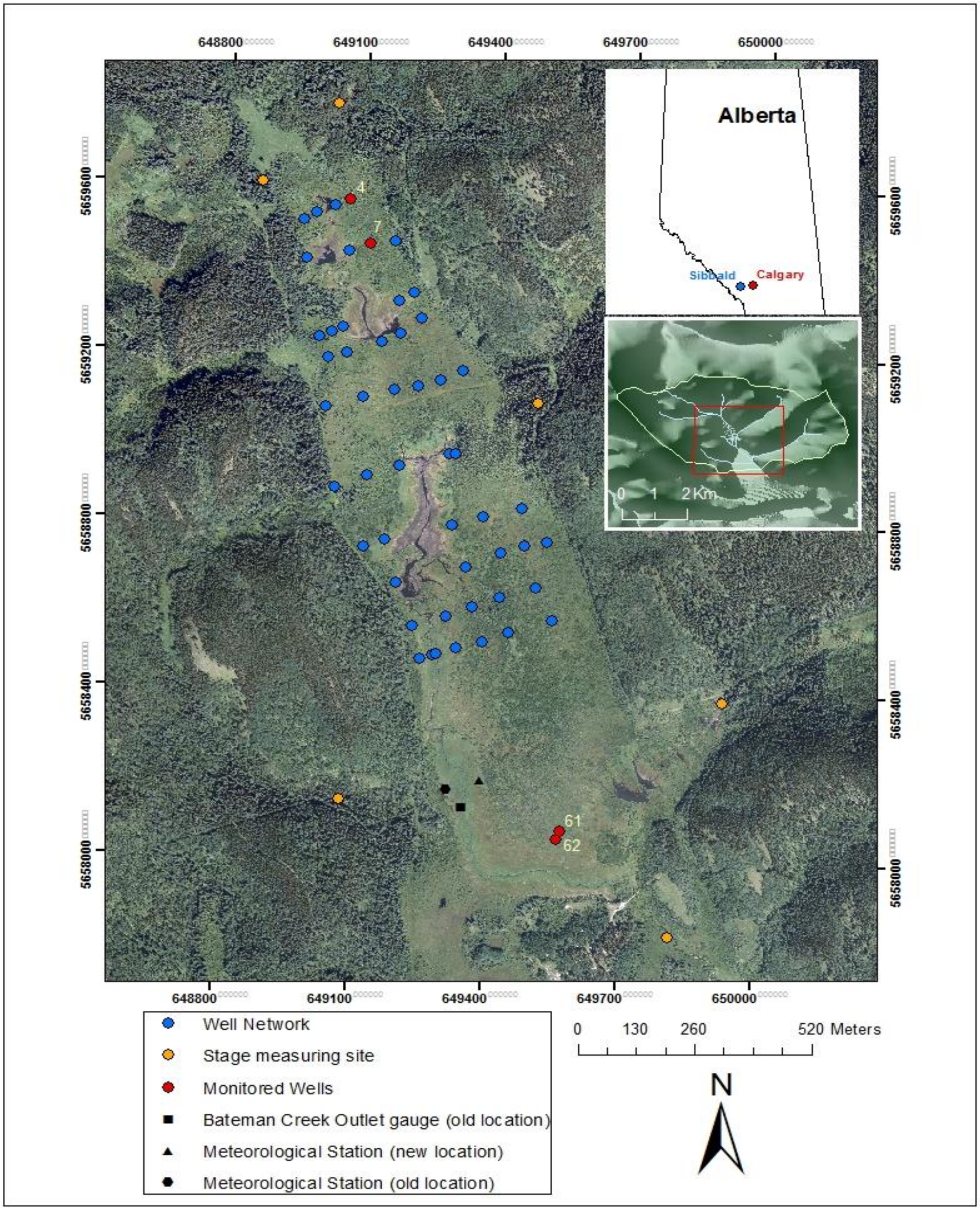


Figure 2.1 Aerial view of the Sibbald Research Wetland with instrumentation. The location of the site within the Province of Alberta and its drainage basin are also displayed. The monitored groundwater wells (4, 7, 61 and 62) are labeled on the map.

2.2 Methods

2.2.1 Components of the water budget

To understand the storage-yield dynamics of the Sibbald Research Wetland, a daily water budget was calculated (mm):

$$\boldsymbol{\Delta S_{obs} = [P + Q_{in} + G_l^{in}] - [ET + Q_{out} + G_l^{out}] \pm G_v^{net}} \qquad (2.1)$$

where ΔS_{obs} is the calculated change in storage, *P* is precipitation in the form of rain and snowfall, Q_{in} and Q_{out} are respectively incoming and outgoing streamflow, G_l^{in} and G_l^{out} are respectively incoming and outgoing lateral groundwater flow, *ET* is actual evapotranspiration, and G_v^{net} is net vertical groundwater exchange between the alluvial aquifer and the peatland. A conceptual representation of the water budget is shown in Figure 2.2. The following sections detail the calculations for each parameter in equation 2.1. The water balance was used to solve for net vertical groundwater exchange (G_v^{net}) between the underlying alluvial aquifer and peat was calculated as the remaining component of the water balance:

$$\boldsymbol{G_v^{net} = \Delta S_{obs} - P - Q_{in} - G_l^{in} + ET + Q_{out} + G_l^{out}} \qquad (2.2)$$

The water balance was calculated for the period 1 June to 13 August 2017. The meteorological station was moved for the research purposes of others to a location 10 m east of its original site (see Figure 2.1) between 6 July and 13 July 2017. This relocation mainly effected ground measurements, with data gaps in sallow and deep volumetric water content, as well as soil heat flux. Thus, to fill in the 6-13 July data gap, relationships were drawn between all three of these variables and water table depths.

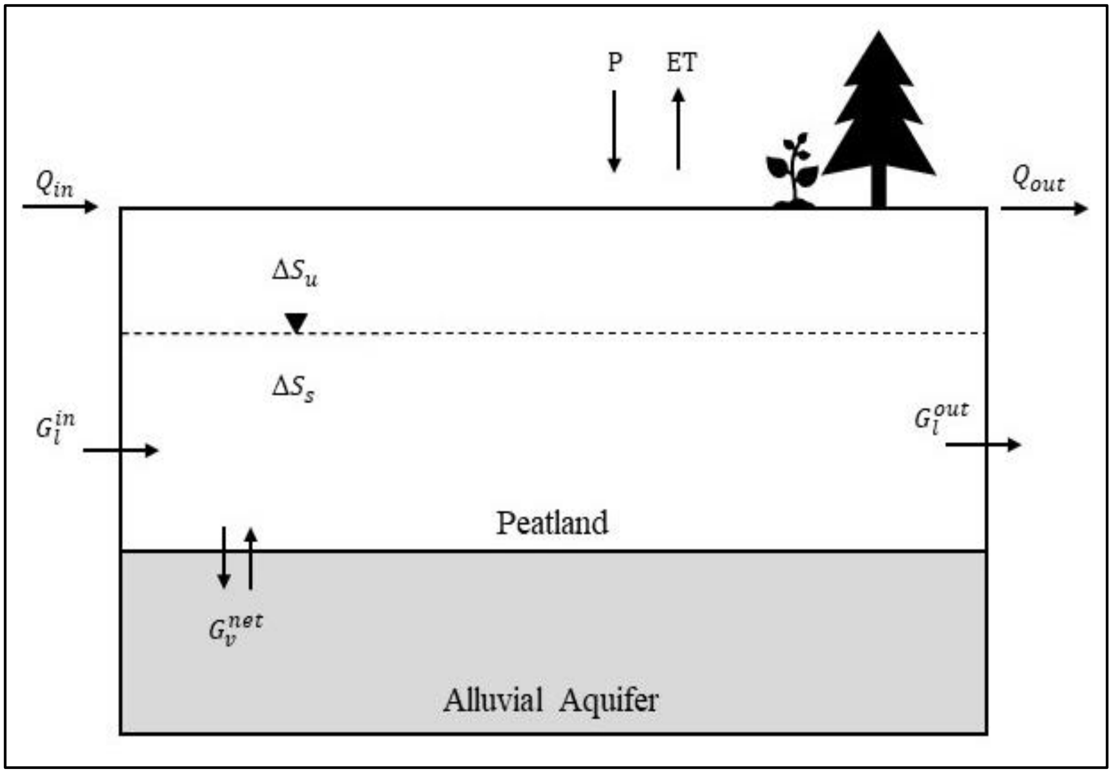


Figure 2.2 Conceptual diagram showing the control volume of the Sibbald Research Wetland. The dotted line delineates the water table. Arrows indicated the direction of each of the water balance fluxes are described in section 2.2.1.

2.2.1.1 *Change in storage*

Observed change in storage (ΔS_{obs}) was found via Spence and Woo (2006) and are illustrated in Figure 2.3:

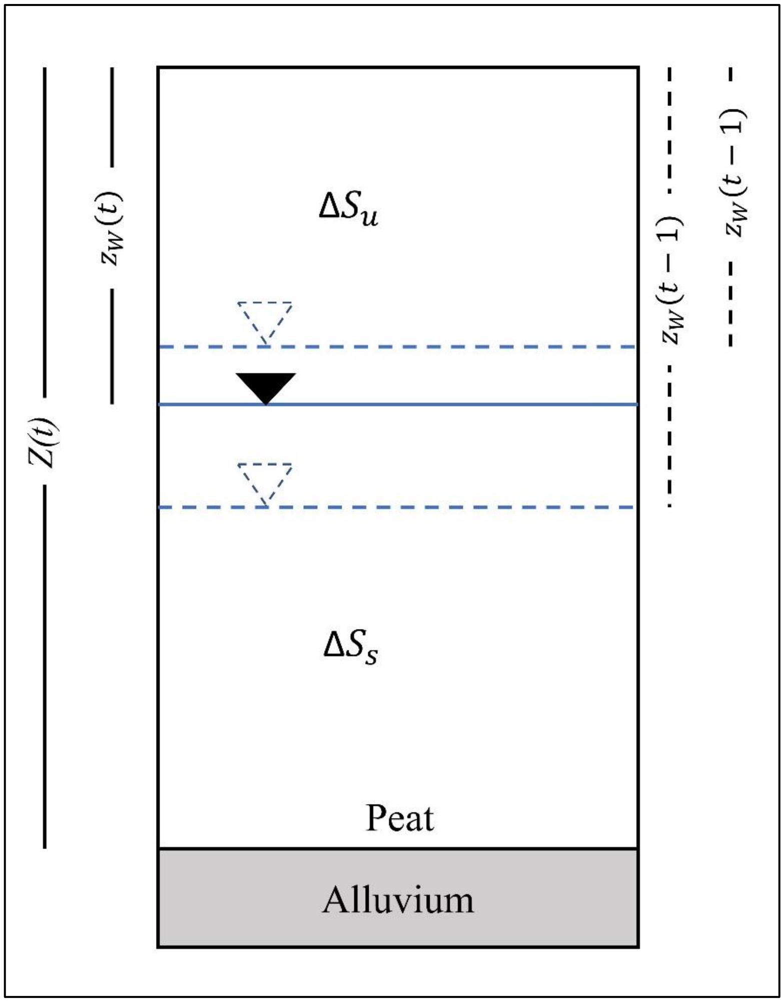


Figure 2.3 Conceptual model used for calculation of change in storage (see equation 2.3 for term definition).

$$\Delta \mathbf{S}_{obs} = \Delta \mathbf{S}_u + \Delta \mathbf{S}_s = \Delta \boldsymbol{\theta}[\mathbf{z}(\mathbf{t}) - \mathbf{z}_w(\mathbf{t})] + \mathbf{S}_y[\mathbf{z}_w(\mathbf{t}) - \mathbf{z}_w(\mathbf{t-1})] \qquad (2.3)$$

where ΔS_{obs} is the total observed change in storage (mm/day), ΔS_u is unsaturated storage (mm/day), ΔS_s is saturated storage (mm/day), $\Delta\theta$ is the change in volumetric soil moisture content (m/m^3) in the unsaturated zone, z(t) is total soil depth (m), z_w(t) is water table depth (m) and S_y is specific yield taken as a value of 0.3 for the entire peat column (Price and Fitzgibbon, 1987), as no site-specific value was available. This value was taken from a boreal peatland with similar plant community composition and peat hydraulic conductivity values. Change in volumetric soil moisture content, Δθ, was calculated using average daily soil moisture data with calibrated soil moisture reflectometers placed horizontally within the peat at 15 and 25 cm depths. Due to the relocation of the meteorological station, a correction factor was applied to volumetric water content data from the new location. Water table depth at the meteorological station was only monitored for the last three weeks of the field season, an average water table depth from wells 4 and 7 (Figure 2.1) was used for this calculation. Water table depth at these two wells was monitored hourly and corrected for barometric pressure. The observed storage change has an expected accuracy of ±25% (Spence and Woo, 2006).

2.2.1.2 *Precipitation*

Daily totals of rainfall (mm) were recorded at the meteorological station. Snowfall was minimal throughout the length of the monitoring season, so snow-water equivalent was not accounted for in water budget calculations. Vegetation interception was not taken into consideration for this measurement due to the difficulty in quantifying it over such as large area with highly variable vegetation cover. Antecedent precipitation index (API), or catchment wetness, was taken as such:

$$\boldsymbol{API_d = P_d + kP_{d-1} + k^2P_{d-2} + k^2P_{d-n}} \quad (2.4)$$

where API_d is the antecedent precipitation index for day *d*, *k* is a decay factor, and P_d is rainfall for day *d*. The decay factor value used was determined from the average decay of the recession limb for the tree largest rain events of the study period, 0.4 (Appendix F, Fedora and Beschta,1989).

2.2.1.3 *Evapotranspiration*

Eddy correlation is the most accurate method for determining evapotranspiration in wetlands (Brummer et al., 2012). However, as eddy correlation was unavailable, actual evapotranspiration (ET) was calculated via the Penman Monteith approach.

To calculate ET, information on the site vegetation cover and characteristics is required. The Sibbald Research Wetland was separated into its three main land-cover types (sedge, willow, open water). Using a maximum likelihood classification computation with 2013 imagery in ArcGIS (ESRI), peatland land-cover was 65% sedge, 23% willow and 12% open water (Appendix B).

Actual ET were calculated separately for sedge and willow using the Penman-Monteith approach (Shuttleworth, 1993):

$$\boldsymbol{AET} = \frac{\boldsymbol{\Delta K_c\left(Q^{*} - Q_g\right) + \frac{\rho_a C_p (e_a^{*} - e_a)}{L_v r_a}}}{\boldsymbol{\Delta + \gamma(1 + \frac{r_c}{r_a})}} \tag{2.5}$$

where K_c is a unit conversion coefficient to provide evaporation in mm/day, Q^{*} is net radiation (W/m²), Q_g is ground heat flux (W/m²), ρ_a is mean air density at constant pressure (kg/m³), C_p is specific heat capacity of air (J/kg/ºC), e_a^{*} is saturation vapour pressure (kPa), e_a is atmospheric water vapour pressure (kPa), L_v is latent heat of vaporisation (ºC), r_a is aerodynamic resistance (s/m), and r_c is canopy resistance (s/m). In this method, aerodynamic resistance was calculated using the method proposed by Oke (1997):

$$\boldsymbol{r_a} = \frac{\mathbf{ln}\frac{(\boldsymbol{z-d})}{\boldsymbol{z_0}}\mathbf{ln}(\frac{\boldsymbol{Z-d}}{\boldsymbol{z_{oh}}})}{\boldsymbol{B^2 u(z)}} \tag{2.6}$$

where *z* is the height of wind measurements (m), *d* is displacement height of vegetation (m), z_o is roughness height of momentum (m), z_{oh} is the roughness height of water vapor (m), *u(z)* is average wind speed at height z (m/s) and B is the von Karman constant. Parameterizations can be found in Table B.1.

Canopy resistance was calculated using the Jarvis approach (Verseghy et al., 1993; Spence et al., 2011):

$$\boldsymbol{r_c = r_s(LAI)(F_s)F_1F_2F_3F_4} \quad \textbf{(2.7)}$$

where r_s is stomatal resistance (r_s= 0sm^{-1}: Lafleur et al. 2005), *LAI* is leaf area index of sedge (0.62: Strack et al., 2006). and willow (2.3: Guan et al., 2010 for alder), F_s is a shelter factor (0.5: Dingman, 2015) and F_1, F_2, F_3, F_4 account for photosynthetically active radiation (W/m²), dependence on vapor pressure deficit (mbar), the influence of soil moisture suction (MPa), and the influence of air temperature (℃) respectively.

Calculation of climatic factors are taken from Verseghy et al. (1993):

$$F_1 = \max(1.0, 500/K_{in} - 1.5) \quad (2.8)$$

$$F_2 = \max(1.0, \mathrm{VD}/5.0) \quad (2.9)$$

$$F_3 = \max(1.03, -\psi/40.0) \quad (2.10)$$

$$F_4 = 1.0 \qquad 40ºC > T_a > 0ºC \quad (2.11)$$

$$= {}^{5000.0}/_{r_c} \qquad T_a \geq 40ºC \ or \ T_a \leq 0ºC$$

where F_3 (parameterized as 1.03 sapric peat from Letts et al., 2000) is the minimum value of soil water suction found for the soil layers contained in the rooting zone (m).

Finally, open water evaporation (E) from beaver ponds was calculated using the Penman equation (Penman, 1948):

$$\boldsymbol{E = \frac{\Delta k_C Q^* + \gamma E_A}{\Delta + \gamma}} \quad (2.12)$$

where E_A is drying power ($E_A = f_u(e^* - e_a)$):

$$f_u = 2.63(1 + 0.537u) \tag{2.13}$$

and u is wind speed at measured height (m/s). Once rates of ET were calculated for each of the landcover types, they were summed based on their areal cover.

2.2.1.4 *Streamflow*

To calculate streamflow, discharge points were added to existing area-velocity rating curves (from 2014) for the north, northeast, east, and west streams flowing into the peatland (Figure 2.1). New gauging stations were established at the Bateman Creek outlet on 26 May owing to the beaver building a dam downstream of the previous gauging station that flooded it (Figure 2.1), and at the southeast stream on 29 May 2017 as well as Bateman outlet. Discharge was measured 1-2 times a week at each stream using a Marsh McBirney flowmeter following Water Survey of Canada guidelines (Lane, 1999). Automated Solinst level loggers were installed in a perforated PVC pipe in each stream and measured stream stage at 15-minute intervals. Levels were corrected for barometric pressure measured with the meteorological station barologger (Solinst ON). Rating curves for the north, northeast, southeast, east and west inlet streams, as well as the Bateman Creek outlet can be found in Appendix D. Runoff ratios were calculated by dividing stormflow by areal precipitation by event (Blume et al., 2007).

2.2.1.5 *Groundwater flow*

Lateral groundwater flow, G_l, was calculated according to Darcy's law:

$$G_l = \frac{K(h_1 - h_2)}{\Delta l} A_c / A_w \tag{2.14}$$

where A_w is the peatland surface area (m²), A_c is cross-section area over which flow is occurring (m²), K is hydraulic conductivity (m/s), h is the height of the water table (taken here as m. a.s.l) and Δl is the distance between the recording monitoring wells. Hydraulic conductivity was estimated in each well using the Hvorselv (1951) method (for K, A_c and A_w see Appendix E).

Water tables from the northernmost transect of the long-term well network (Westbrook and Bedard-Haughn, 2016) were used to estimate inflowing and outflowing G_l. Level loggers recorded water table elevation in wells 4 and 7 for incoming lateral flow. Two new monitoring wells, wells 61 and 62, were installed toward the outlet of the peatland, given the downstream relocation of the outflow stream gauge, and were also equipped with level loggers on July 5, 2017 after the ground frost thawed.

2.2.2 Evaluating frozen ground

Frozen ground was monitored every 1-2 weeks during the thaw period. Frost was monitored at approximately the same location at each well of the well network. A graduated steel rod (183 cm) was pushed into the ground until resistance to further penetration occurred, and the depth recorded. (Woo and Xia, 1996). If the rod could be pushed into the peat near, or beyond its capacity, the frost table was assumed thawed. Concrete frost could have been occasionally encountered instead of ground ice. Once collected, frost table observations were spatially interpolated in ArcGIS using kriging for each observation date.

2.2.3 Peatland hydrological function

Many studies have shown that hydrological elements, including wetlands, have two main hydrological functions: storage and discharge (Roulet and Woo, 1986; Black, 1997; Hayashi, et al., 2016). Hydrological units store water when their change in storage (ΔS) is greater than total runoff (R). Inversely, discharge occurs when total runoff is greater than the change in storage. Spence (2007) further divides discharge into two additional functions: transmission and contribution. A wetland is transmitting water to an adjacent landscape unit when internally generated runoff (IGR), or outflow – inflow, is greater than inflow, and contributing water when internally generated runoff is smaller or equal to inflow (Figure 2.2). To account for additional storage loss, storage was further divided into evapotranspiration and peatland recharge. ET occurs with a negative change in storage, while peatland recharge occurs when the change in storage is positive. Peatland hydrological functions were determined for each day of the study period by looking at all the parameters of the water balance.

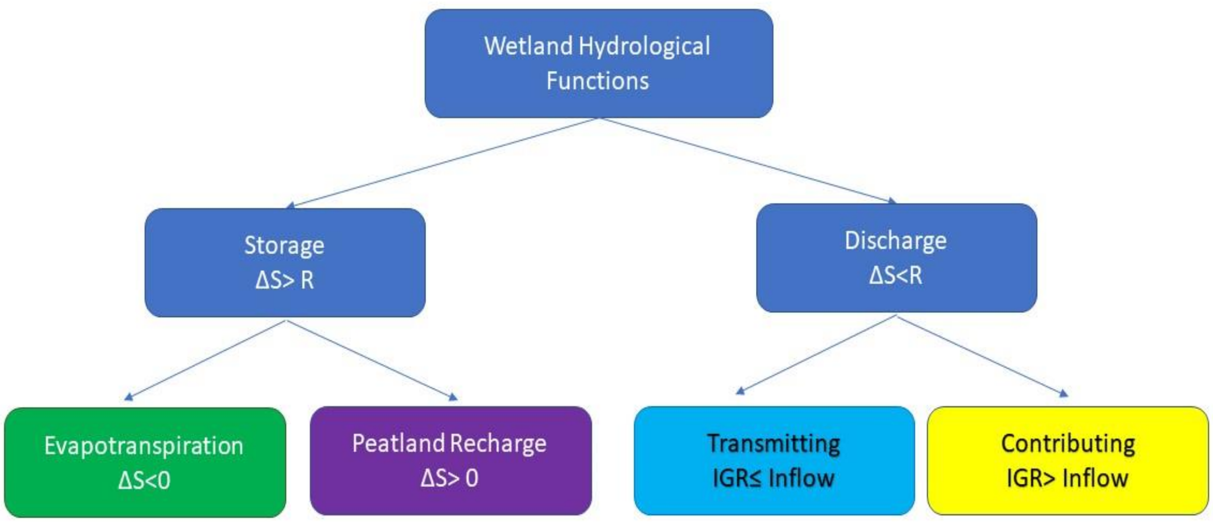


Figure 2.4 Flow chart representing Spence's (2007) definition of wetland hydrological functions with an addition of an evapotranspiration function. Note that internally generated runoff (IGR) is the difference between outflow and inflow.

CHAPTER 3 - RESULTS

3.1 Frost

Frost table surveys indicate that the peatland first started to thaw in the southern end of the well network (Figure 3.1). Thawing gradually progressed northward throughout the spring and summer. Frost tables were shallower than 50 cm below the surface as late as 28 June along the five northernmost transects. In a few of the northern wells, frost persisted until 28 July. Plots of time of frost table thaw against peat depth (Appendix A) shows frost table was shallowest in places where peat depth was shallowest (most notably in the north end of the peatland). Similarly, frost table was deepest in areas where the peat was deepest (at the southern extent of the well network).

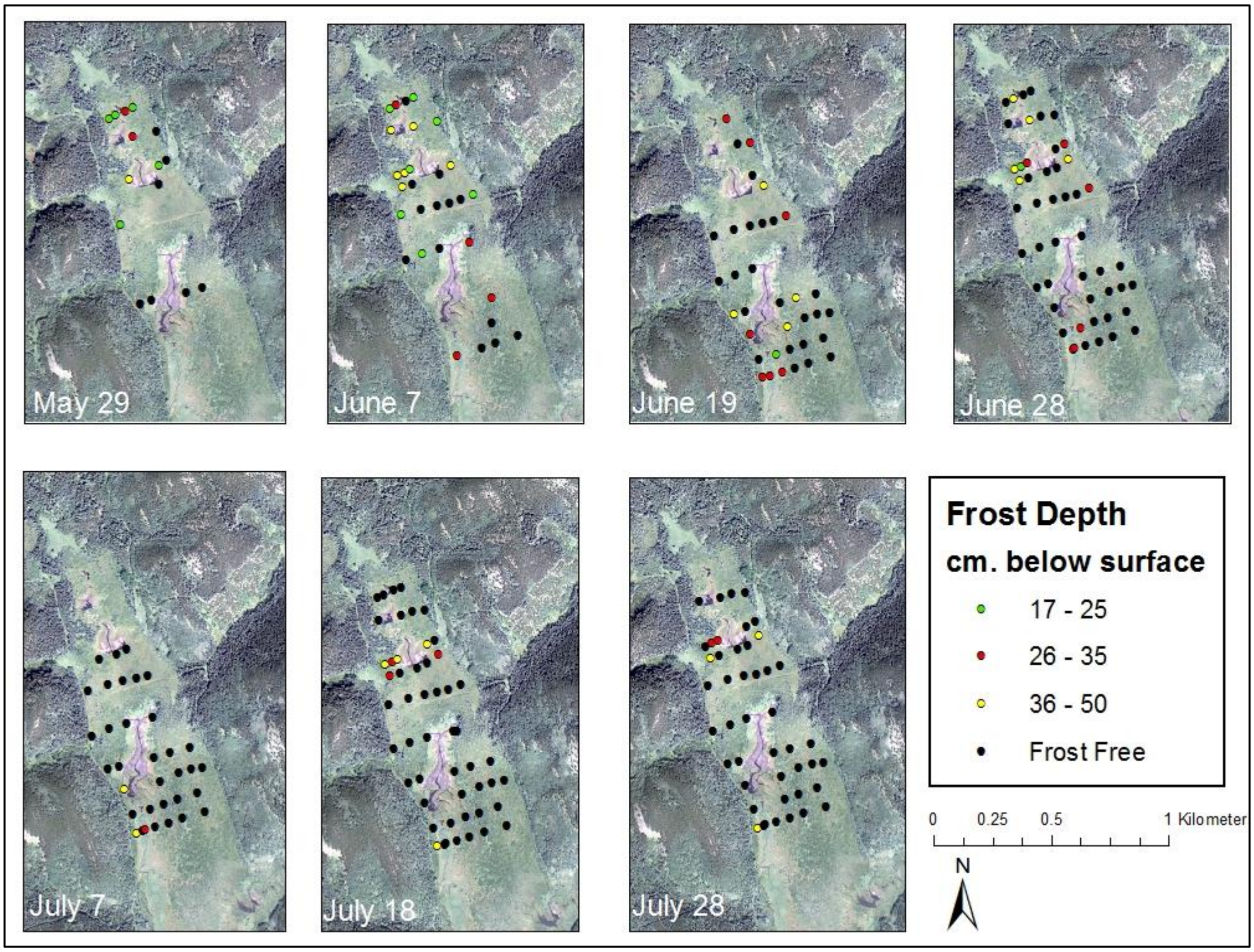


Figure 3.1 The temporal distribution of frost throughout the months of May, June and July 2017 at the Sibbald Research Wetland. The dots shown in each map represent the areas in which frost measurements were taken on each day of the survey.

3.2 Water Table Dynamics

Water tables at all four wells (W4 and W7 in the north end and W61 and W62 in the south end) showed similar trends (Figure 3.2): they were highest in spring, and dropped as the summer progressed. Additionally, water tables rose during precipitation events, and dropped during periods of drought. However, water tables at W4 and W7 were on average 0.31 m lower below the surface than water tables at W61 and W62 (where average water table at W4 and W7 was 0.58 m, and 0.27 m below the surface for W61 and W62). Of note, wells 61 and 62 were 75 m and 60 m away, respectively, from a beaver dam at the outlet of Bateman Creek, and so may have been influenced by the dam (Karran et al., 2018). From 28 June to the beginning of August, well 4 and 7 water tables declined at an average rate of 13 mm/day, and well 62 and 62 water tables declined at a rate of 5 mm/day. Rainstorms on 10-11 July (32.5 mm) and 23 July (3.4 mm) caused transient water table rises. During the first storm, W4 and W7 water tables rose from 0.38 m and 0.59 m below the surface on 10 July, to a depth of 20.5 m below the surface on 11 July. Similarly, W61 rose from 0.20 m to 0.11 m below the surface, while W62 rose from 0.18 m to 0.08 m below the surface. Finally, water tables at all four wells showed clear diurnal patterns early in the monitoring period, a result of evapotranspiration. However, once W7 dropped <0.93 m on 3 August, these fluctuations were no longer visible. Coincident with this, the rate of water table decline at W7 slowed to 9 mm/day.

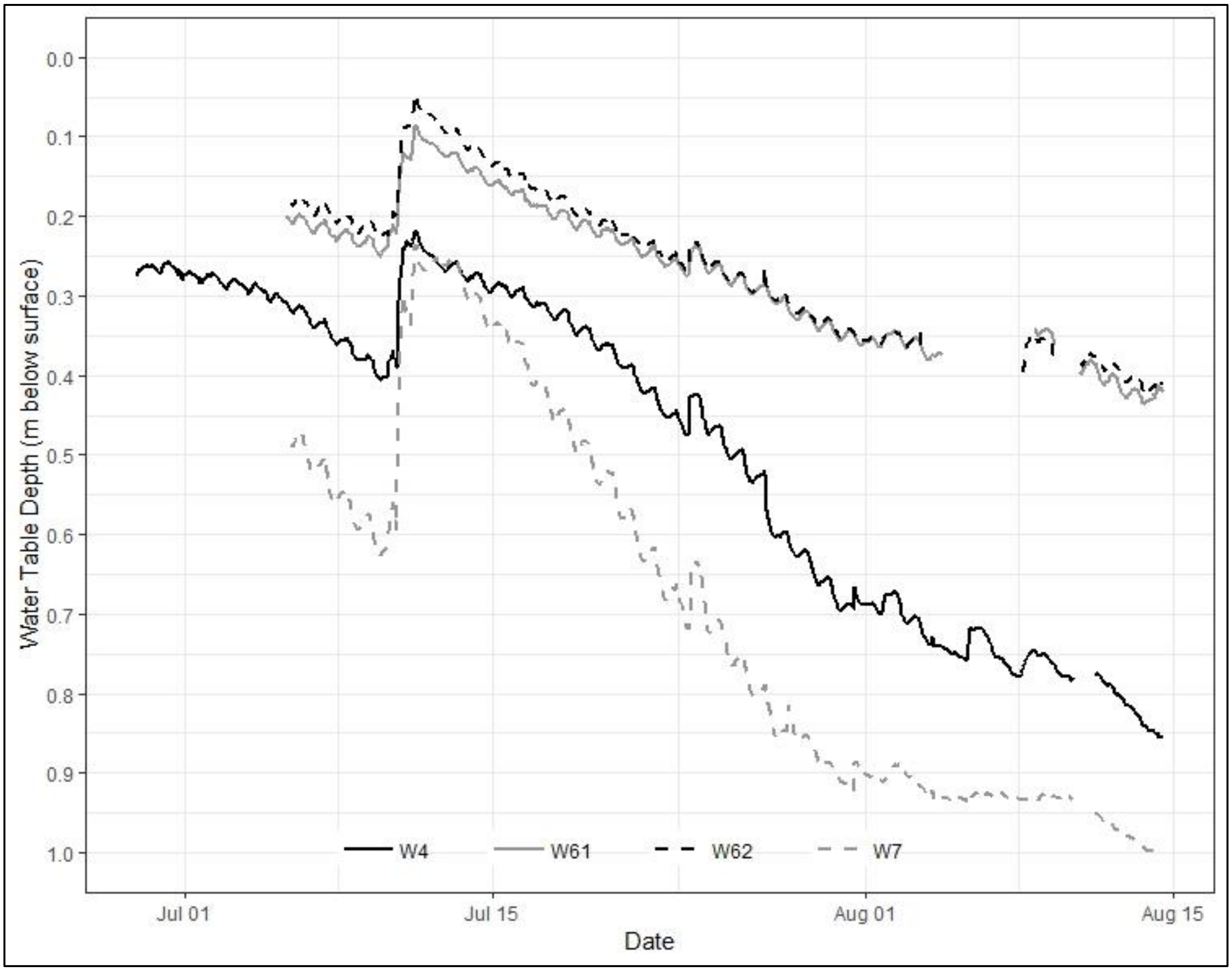


Figure 3.2 Water table depths below the peat surface (m) at the Sibbald Research Wetland, 1 June-13 August.

3.3 Soil Moisture Dynamics

Observations from both shallow and deep volumetric water content (VWC) measured at 25 cm and 50 cm depths were impacted by the relocation of the meteorological station (Figure 3.3). Both shallow and deep peat at the original location of the meteorological station contained less water than the peat in the new location. From 1 June to 5 July, peat at 25 cm below the surface varied with precipitation: large rainstorms caused large increases in shallow soil moisture, and small precipitation events caused marginal increases in shallow moisture. On 11 June, during a rainstorm that produced 19.1 mm of rain, shallow VWC reached 76%. Additionally, shallow VWC decreased in between rainstorms at an average rate of 1.5%/day. On July 14, after relocation, shallow VWC shifted to 78%, as there was quite a large rainstorm (July 10-11) during the week that VWC was not monitored. Small rain events (<5 mm/day) caused little variation in soil moisture, which continued to drop at an average rate of 0.6%/day until the end of the study period.

Soil moisture at 50 cm below the surface ranged between 76% and 74% prior to the relocation of the meteorological station. Precipitation events caused little variation in deep VWC which continually decreased during this time. On July 6, deep VWC was at 74% and increased to 76% during the 10-11 July rainstorm which produced 32.5 mm of rain. Subsequently, deep VWC decreased until the end of the study period, with minimal reaction to the small precipitation events (<5 mm) during this time.

Figure 3.3 Daily precipitation, shallow and deep volumetric water content, corrected for the relocation of the meteorological station at the Sibbald Research Wetland, 1 May – 13 August, 2017

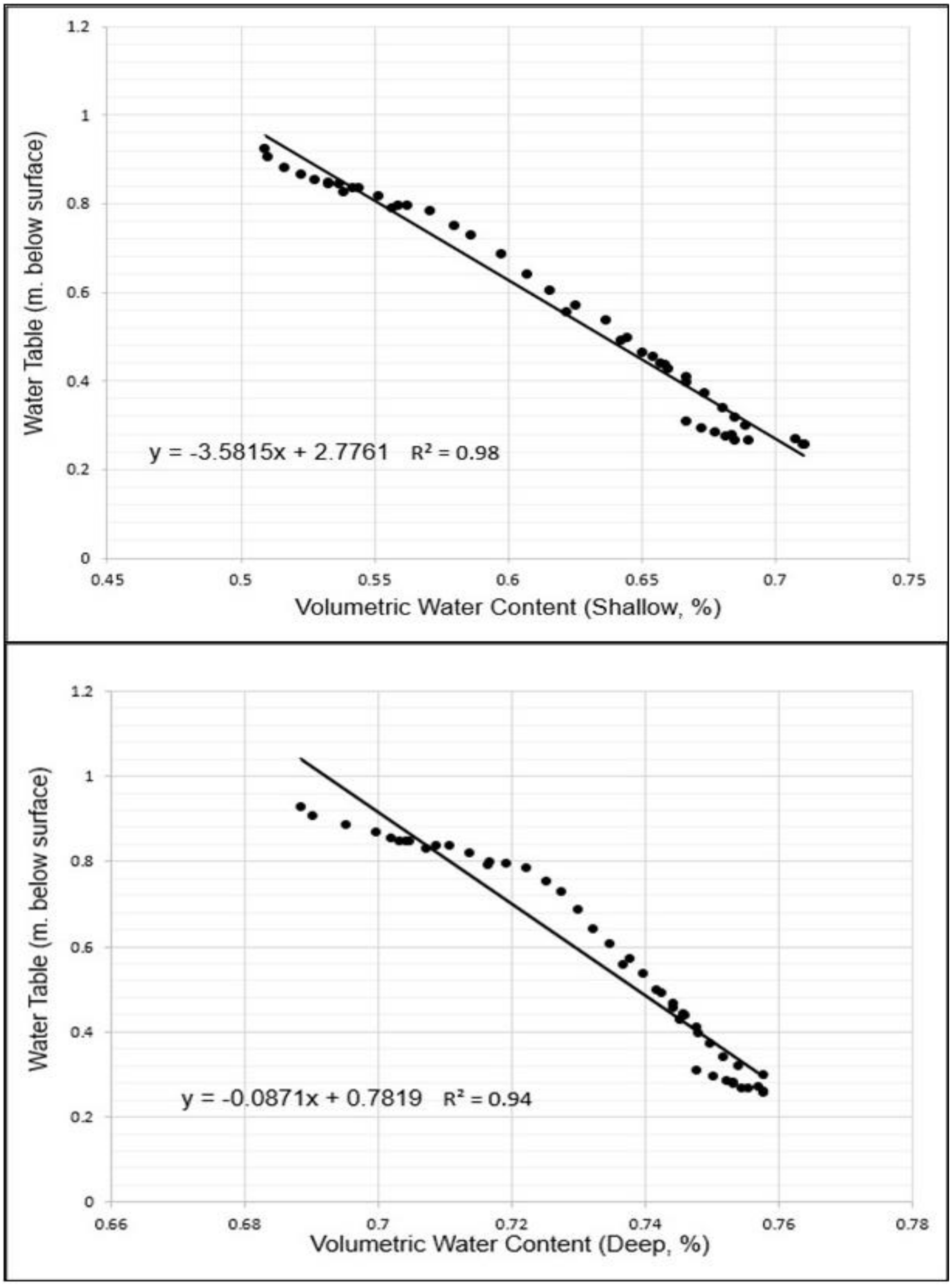


Figure 3.4 Shallow and deep volumetric water content (%) vs. water table depth (m. below surface). Strong linear correlation between the variables is indicated by the R^2 value.

A strong positive linear relationship between water table and both shallow (Figure 3.4, R^2= 0.98) and deep (Figure 3.4, R^2= 0.94) volumetric water content was observed. As VWC increased, so did water table height. This occurred both before and after the meteorological station was moved.

3.4 The Water Balance

3.4.1 Precipitation

Rainfall contributed 118 mm of water to the peatland during the study (Figure 3.5 and 5.13); each month had less rain than the last. Thus, June, July and the first two weeks of August saw 65.48, 40.2, and 11.9 mm of rain, respectively. A greater number of days with rain occurred in June, with >0.5 mm of rainfall occurring 2-4 times a week. Two of the largest precipitation events of the study period also occurred in June: 8-10 June (33.3 mm) and 13-14 June (18.3 mm). Although the second largest precipitation event of the study period occurred 10-11 July (32.5 mm), this month was mainly characterized by long periods without rainfall. For instance, no precipitation events occurred between 2-9 July and 14-22 July. Furthermore, even though August was the month in the study period with the least amount of rainfall, it rained 2-4 days each week. However, these events were small (<3 mm/day). Insignificant snowfall occurred during the first few days of the study, amounting to <0.5 mm, so snow was not further considered.

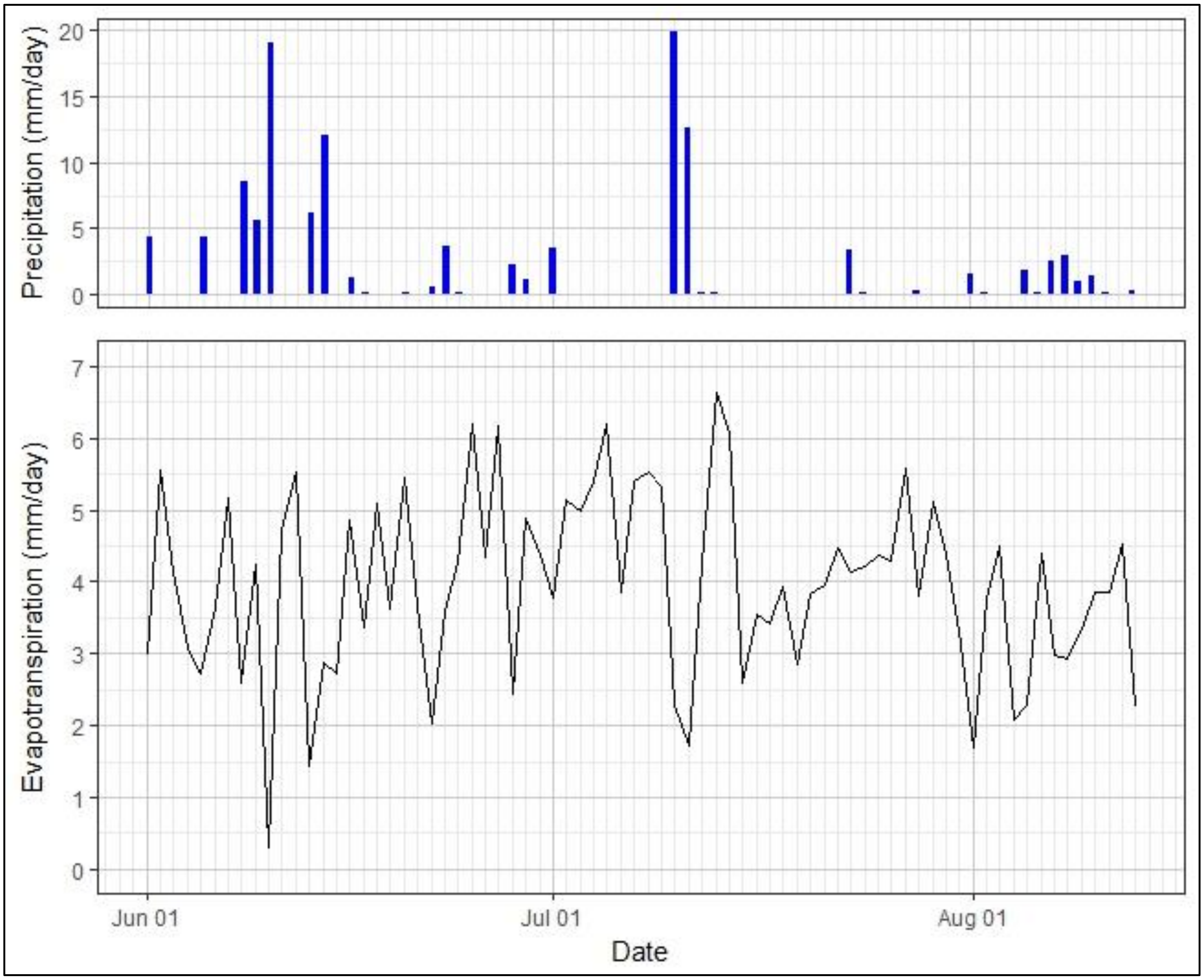


Figure 3.5 Daily precipitation and evapotranspiration at the Sibbald Research Wetland, June-August 2017.

3.4.2 Evapotranspiration

Evapotranspiration accounted for the second largest loss of water from the peatland after outgoing streamflow (Figure 3.5 and 3.13). By the end of the monitoring period 290 mm of water was lost to the atmosphere. Daily mean ET was 4.0 mm/day. ET was highest between mid-June and early July, with an average rate of 4.4 mm/day, and gradually started to decrease thereafter as net and incoming shortwave radiation decreased (Appendix C). ET was lowest on days where there were large amounts of rainfall. On 10 June ET was recorded as 0.3 mm/day. This value is likely due to the net radiometer being covered in water during a large rainstorm.

Diurnal fluctuations captured in the water table observations (Figure 3.2) indicate that aerodynamic and canopy resistance played an important role in regulating water tables. Indeed, the diurnal fluctuations in Figure 3.2 suggests that phreatophytes, the dominant plant species at the Sibbald Research Wetland, had full access to subsurface water throughout most of the summer. Dominant phreatophyte communities at the Sibbald Research Wetland include willows and sedges (Appendix B). Only during the last 10 days of the study do water tables in the north end (well W7) drop below the rooting zone, indicated by the cessation of a diurnal fluctuations in water table.

3.4.3 Lateral groundwater flux

Incoming (G_l^{in}) and outgoing (G_l^{out}) lateral groundwater flux are seen in Figure 3.10. G_l^{in} was higher than G_l^{out} for the entire monitoring season, with values ranging between 0.079-0.087 mm and 0.00027-0.00073 mm/day, respectively. They also showed similar trends: from 5-9 July, both $G_l^{in}$and G_l^{out} increased, with a sudden drop on 11 July during the 10-11 July rainstorm. On 11 July, a flow reversal occurred, where G_l^{out} was reduced to -0.00027 mm/day, while G_l^{in} flows hovered close to 0 mm/day on 12-13 July, when groundwater reached 0.079 mm/day on both days. Both $G_l^{in}$and G_l^{out} both took about two weeks to recover from this reduced flow rate, with G_l^{in} and G_l^{out} peaking on July 26 (0.087 mm/day) and August 2 (0.00073 mm/day), respectively. After this, groundwater flux decreased at both the inlet and outlet for the rest of the season. Extremely small G_l^{out} fluxes may be attributed to the proximity of W61 and W62 to the beaver dam at the outlet of Bateman Creek. Since these wells are situated within 150 meters of the beaver dam, it is likely that water tables within this zone were somewhat stabilized (Karran et. al., 2018).

Cumulative net lateral groundwater flux was negligible as total cumulative G_l^{in} was 3 mm while cumulative G_l^{out} was 0 mm over the monitoring period (Figure 3.13).

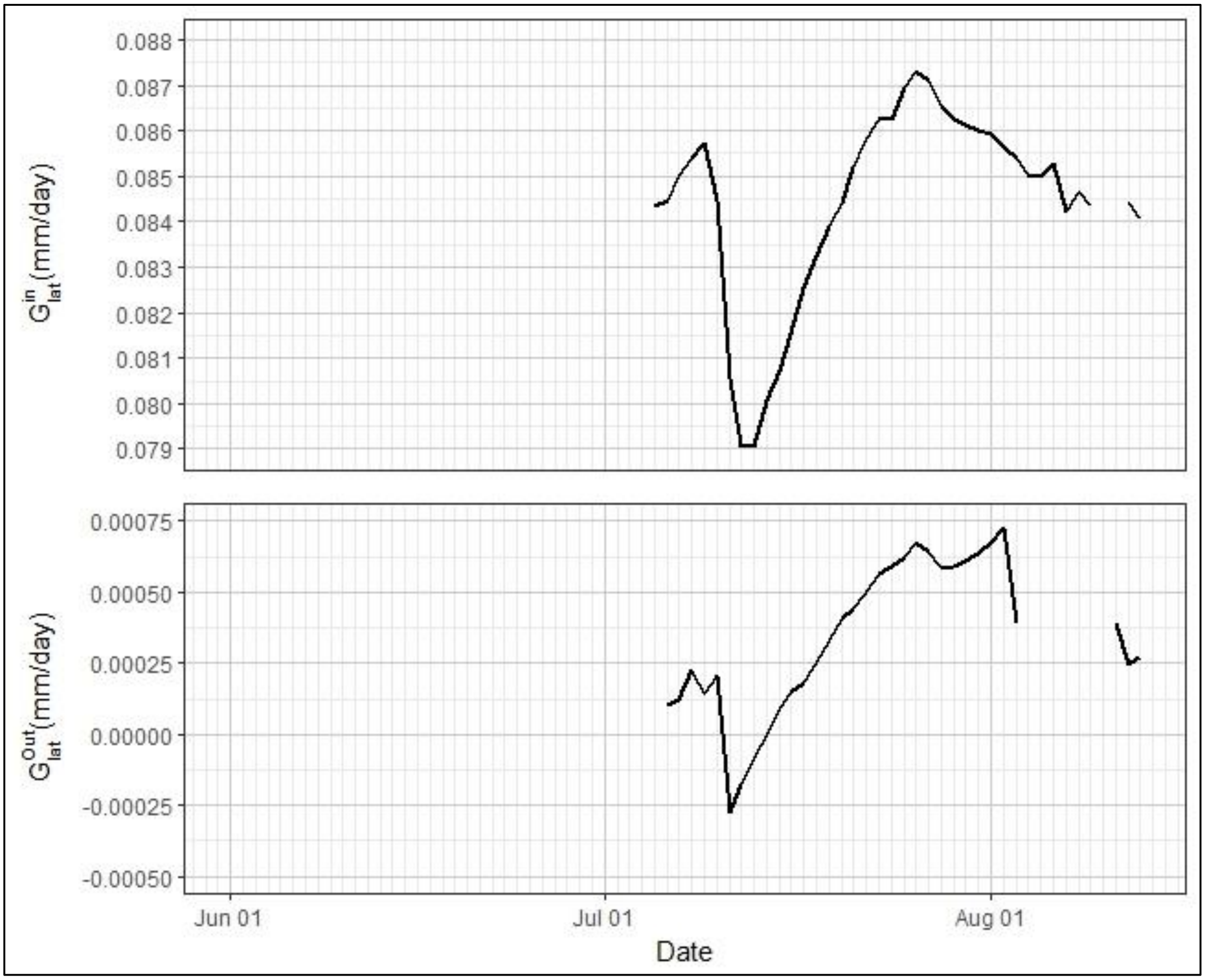


Figure 3.6 Incoming (G_{lat}^{in}) and outgoing (G_{lat}^{out}) groundwater flow from the Sibbald Research Wetland, from 5 July- 13 August 2017. Soils were frozen before 5 July and so measurement of lateral groundwater flow was not possible.

3.4.4 Streamflow

Hydrographs for the southeast, east, northeast, north and west inlet streams (Figure 3.6) show decreasing flow during the length of the monitoring period. The southeast stream discharged the most water during this time, with 50 mm of water flowing into the peatland. This was followed by the east, north, west, and northeast streams, which contributed 21, 19, 10 and 4 mm of water respectively (Table 3.1). Discharge in the southeast stream stayed relatively constant, ranging between 0.6-0.9 mm/day, with discharge at the end of the period being very similar to that of the beginning of the period. Long-term observations indicate that this is likely due to several springs contributing constant flow to the stream through the spring and summer (*C. Westbrook, personal observation).* The southeast and east streams also had the highest discharge at the start of the monitoring period (flowing at rates of 0.6 and 0.5 mm/day, respectively). All streams had increased discharge during the two large rainstorms in June, and all but the west and southeast streams showed an increase in discharge during the large precipitation event in July. The west and northeast streams responded to the small amount of rainfall (<5 mm) received during the last seven days of the monitoring season. The first stream to reach baseflow was the west stream (21 June), which hovered around 0.1 mm/day for almost the entire study period, until August 8, when discharge peaked at 0.27 mm/day during a series of small precipitation events. All other streams reached baseflow nearly one month after this (July 18). The north stream was the only stream to run dry.

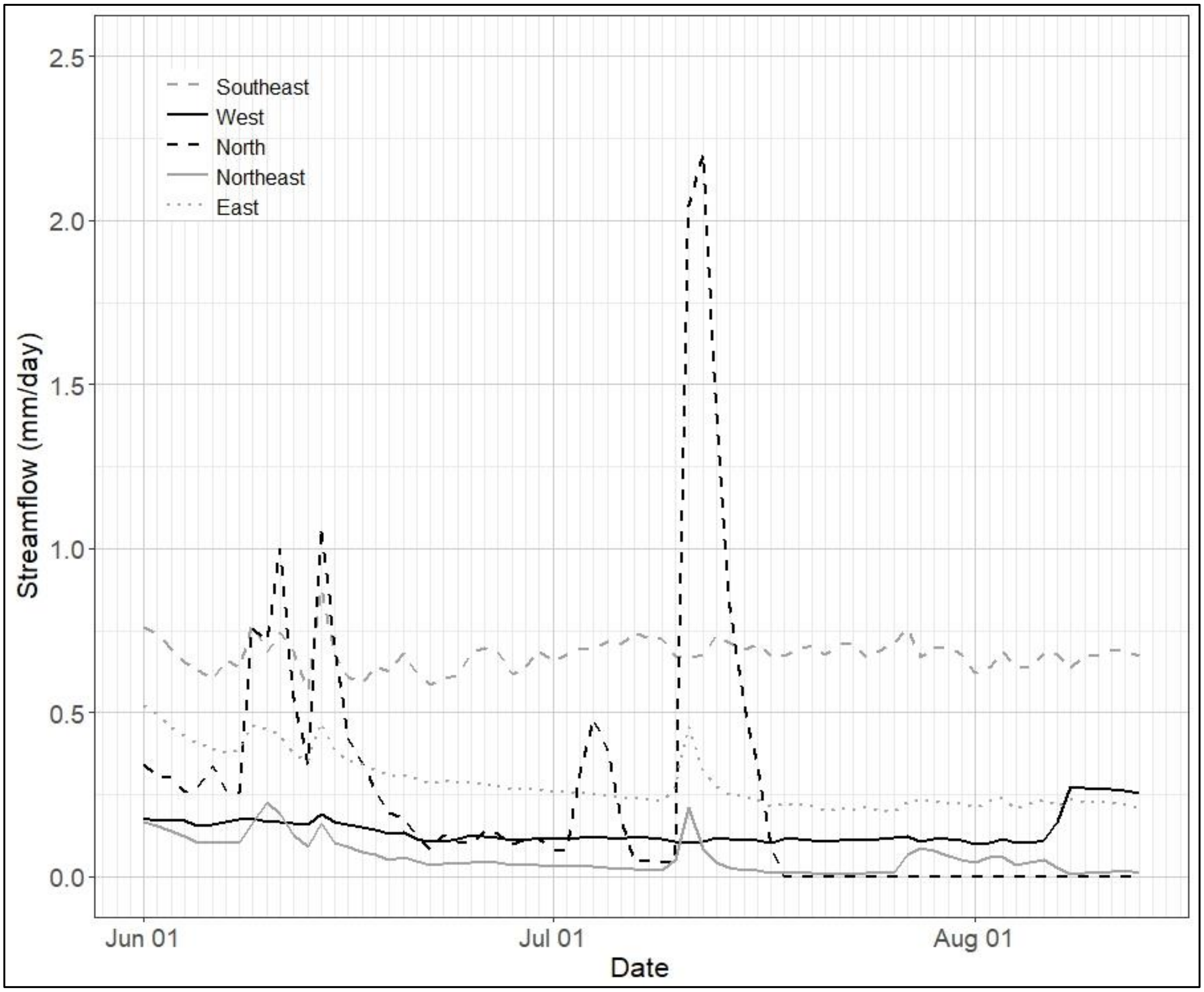


Figure 3.7 Total cumulative flows for all the inlet and outlet streams at the Sibbald Research Wetland from 1 June-13 August.

Table 3.1 Daily stream flow of the Southeast, West, North, Northeast, and East inlet streams at the Sibbald Research Wetland, 1 June-13 August 2017.

Stream	**Southeast**	**East**	**Northeast**	**North**	**West**	**Total Inflow**	**Bateman Creek Outlet**
Total cumulative flow (mm)	50	21	4	19	10	104	301

Outgoing streamflow from Bateman Creek, Q_{out}, had high flows towards the beginning of the study period, and reached baseflow the last week of July (Figure 3.7). On 1 June, streamflow at the outlet was 7.5 mm/day and reached baseflow on 21 July (1.6 mm/day). During the 9-11 and 14-15 June rainstorms, Q_{out} increased to 11.2 and 11.6 mm/day, respectively. During the large rain event of 10-11 July, outflow increased from 2.3 to 7.3 mm/day. This was the last rain event that produced any major peaks in the outlet hydrograph; when the outlet reached baseflow, small rain events in August (<5 mm) generated little reaction from the stream.

During the three-month study period, total incoming (Q_{in}) and outgoing (Q_{out}) streamflow hydrographs differed. Q_{out} was always higher than Q_{in} (Figure 3.7). Streamflow always responded to major precipitation events but streamflow out of the peatland increased during most large (>10 mm) and small (<5 mm) precipitation events by 2-4 times more than that of incoming streamflow. Q_{in} fluctuated little during rain events, even the largest ones, and stayed close to baseflow (~1 mm/day) for most of the study period. Total cumulative discharge for Q_{in} was 103 mm, whilst cumulative discharge for Q_{out} was 301 mm for entire study period (Figure 3.13).

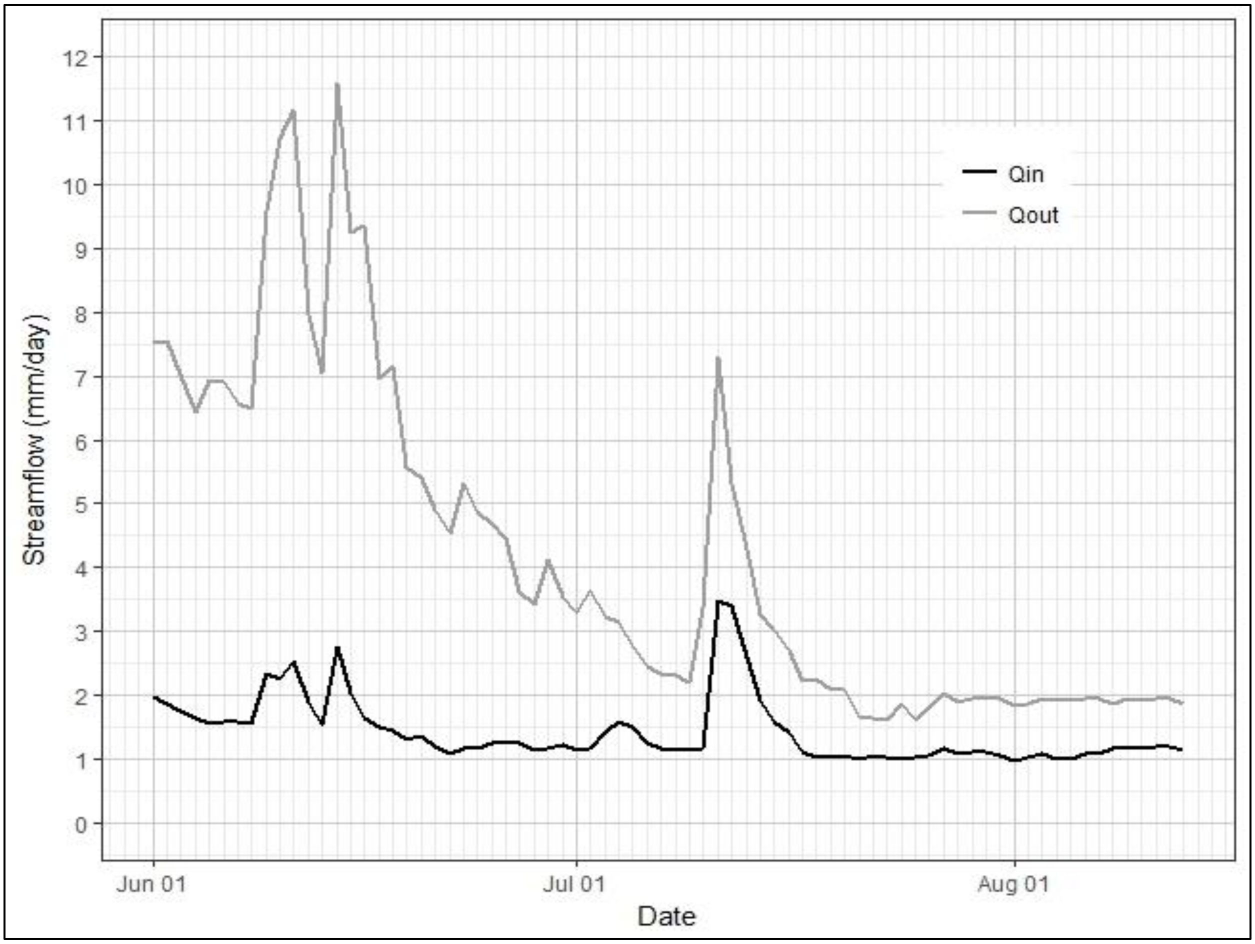


Figure 3.8 Total inflow (Qin) from all inlet streams and outflow (Qout) from Bateman Creek Outlet at the Sibbald Research Wetland, 1 June-13 August 2017. Streamflow was scaled to the area of the peatland (0.71 km²).

Antecedent precipitation index (API, Figure 3.9) and event-based runoff ratios for storms with >5 mm rainfall (Figure 3.10) were plotted with the separated hydrograph at Bateman Outlet. They were highest in June and during large rain events. During periods of high API, event-based runoff was highest and vice versa. Runoff ratios, calculated from stormflow generated during the length of the storm and only for storms greater than 5.0 mm, were highest in spring. Additionally, the 13-14 June rainstorm produced a higher runoff ratio than the 8-10 June rainstorm. Virtually all precipitation during the 10-11 July storm did not reach the stream outlet.

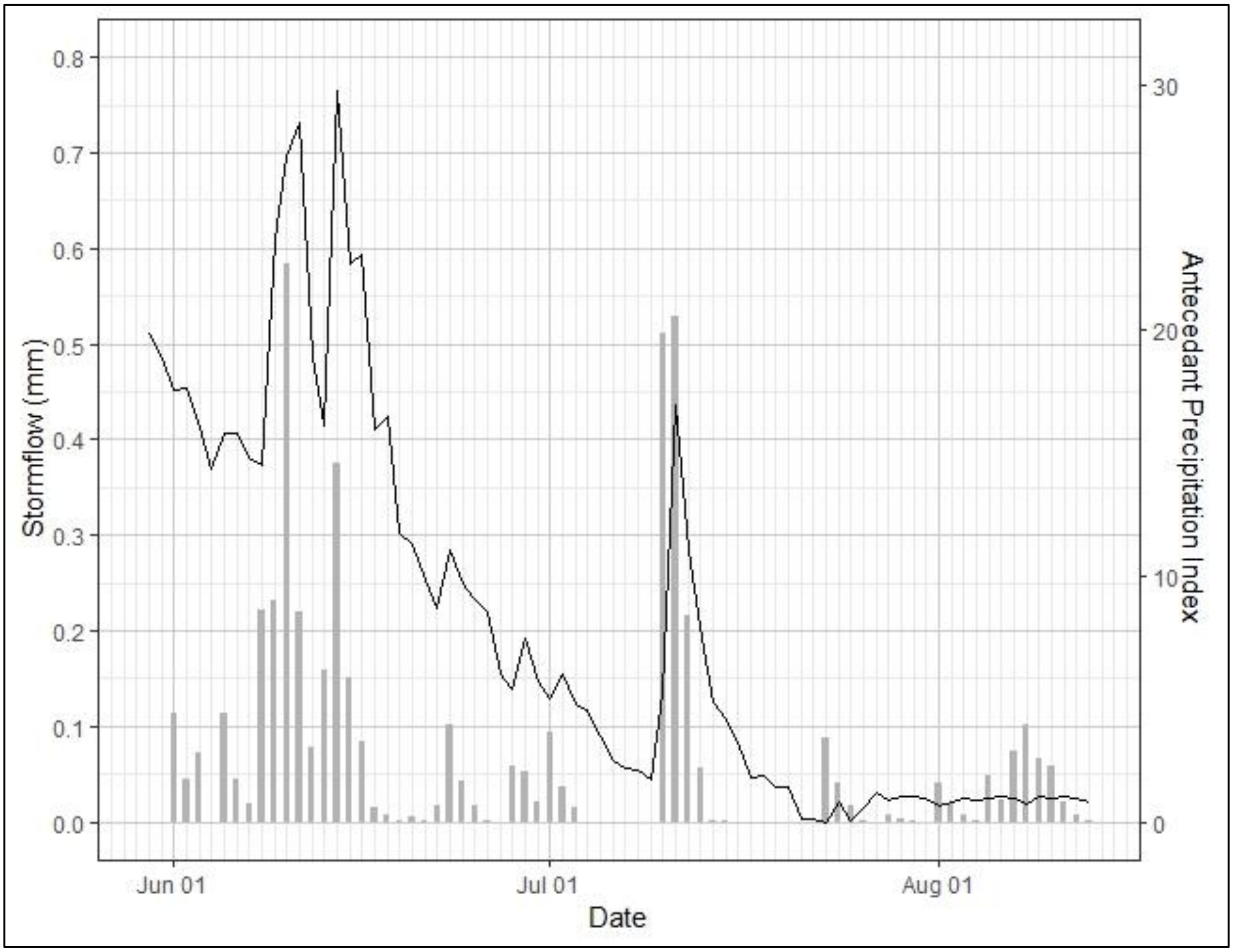


Figure 3.9 Separated hydrograph of Bateman Outlet and antecedent precipitation index of the Sibbald Research Wetland (0.71km²), 1 June- 13 August, 2017.

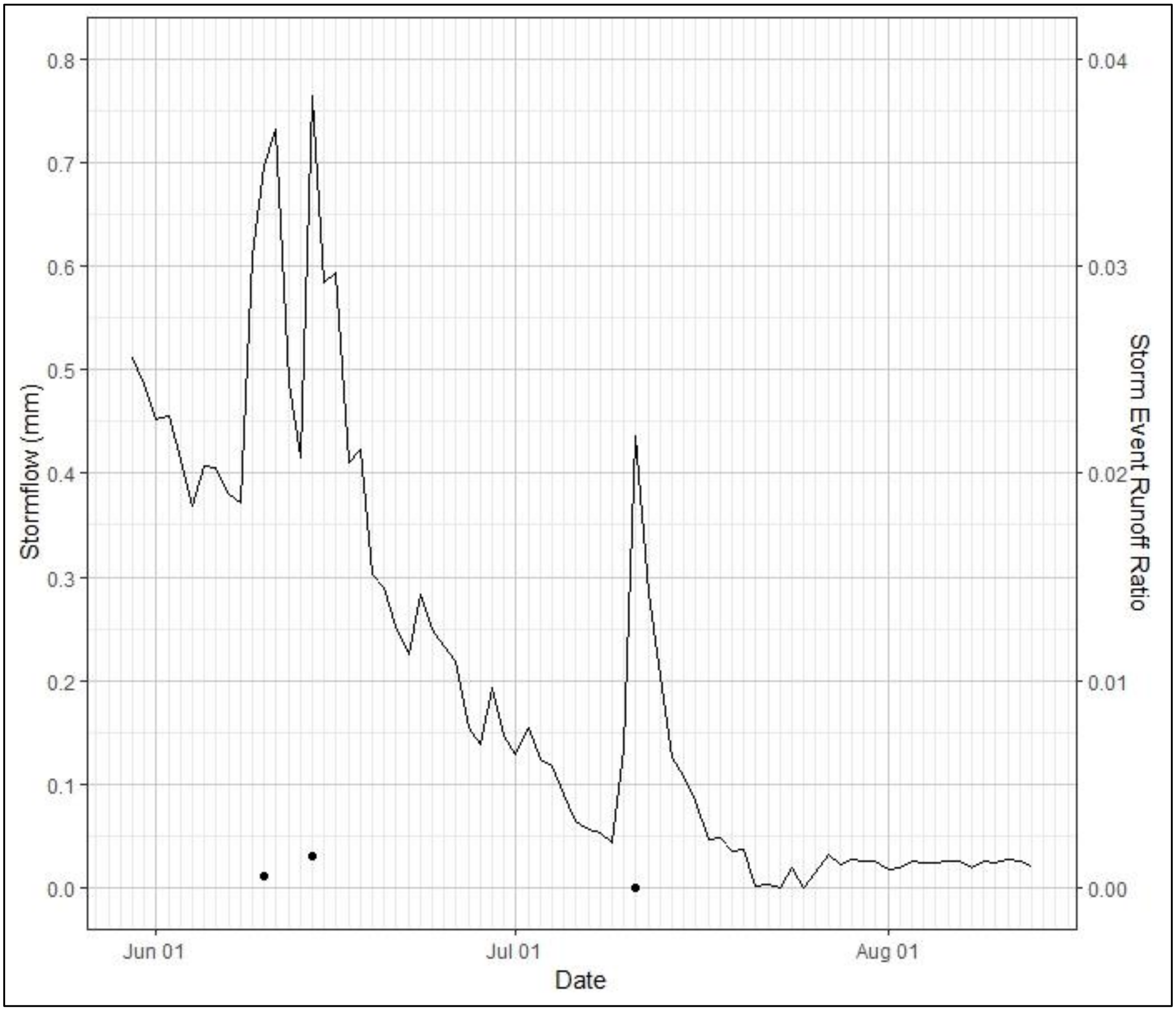


Figure 3.10 Separated hydrograph of Bateman Outlet and event-based runoff ratios of the Sibbald Research Wetland (0.71km²), 1 June- 13 August, 2017. Runoff ratios are represented by points of the graph and were only calculated for storms where precipitation was greater than 5mm of water.

3.4.5 Daily observed change in storage

Figure 3.11 shows change in daily observed storage (ΔS_{obs}, mm/day) as a function of precipitation (mm/day). Generally, the peatland gained the most water in June, and lost water in July and August. On average, the peatland lost 4.5 mm of water each day. The peatland tended to gain water during all precipitation events, and lost water in between rainstorms. The largest fluctuations of observed storage occurred at the beginning of the study period during the 9-11 June and 14-15 June rainstorms. Between 8-11 June, the peatland went from losing 3.0 mm to gaining 121 mm of water. During the two dry days immediately following this rain event, the peatland lost 64 mm of water. Similarly, the rain on June 14 allowed the peatland to gain another 44 mm of water, followed by another large loss of 35 mm of water, during another dry period on June 16. The peatland also gained 39 mm of water during the large precipitation event of July 11. These extremes in storage may be attributed to a relatively large error in the water budget as the calculation does not account for surface water being stored by the beaver ponds as well as the error from all other calculations. Finally, the peatland ran a water deficit over the monitoring period, where cumulative ΔS_{obs} was -330 mm (Figure 3.13).

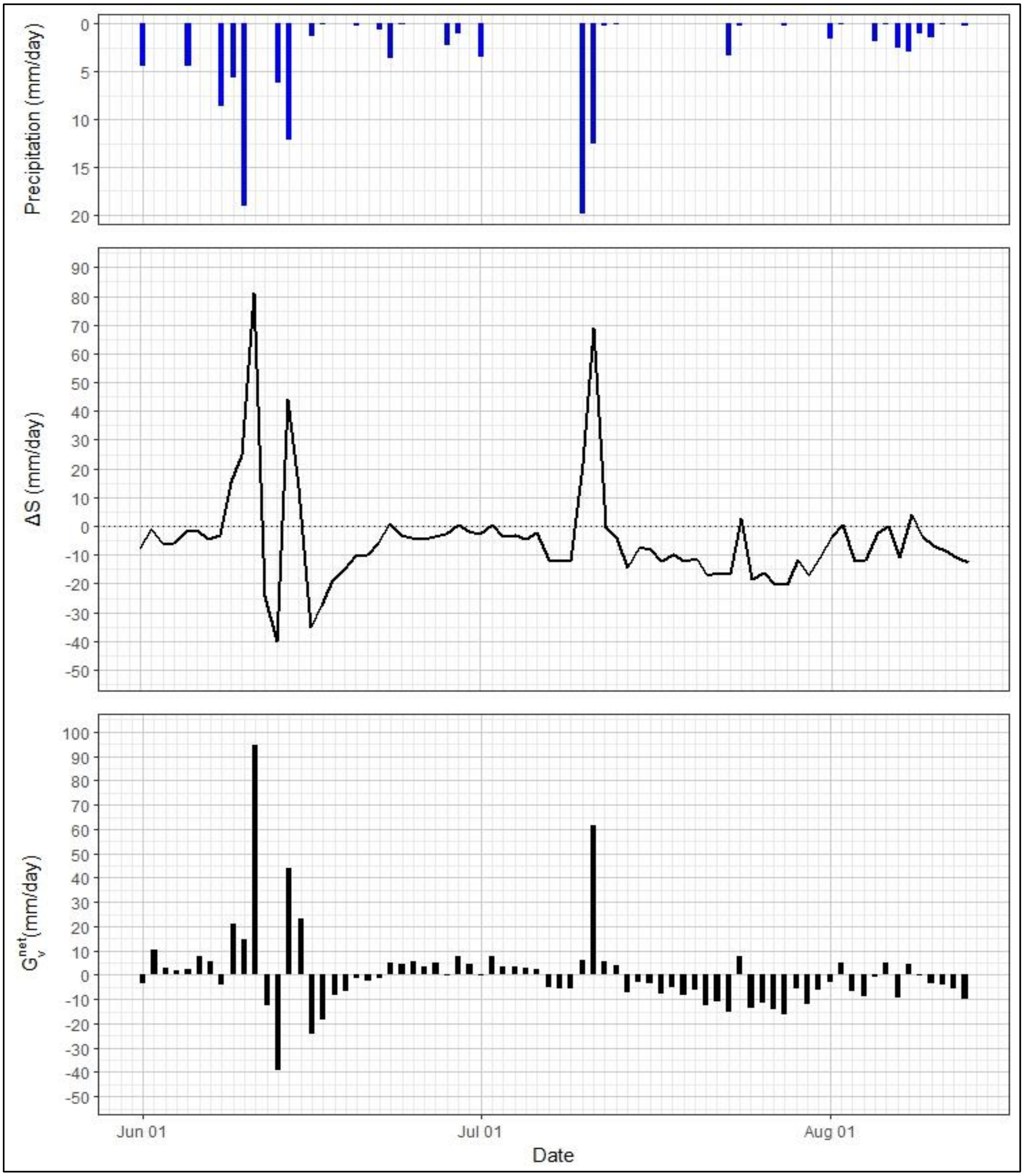
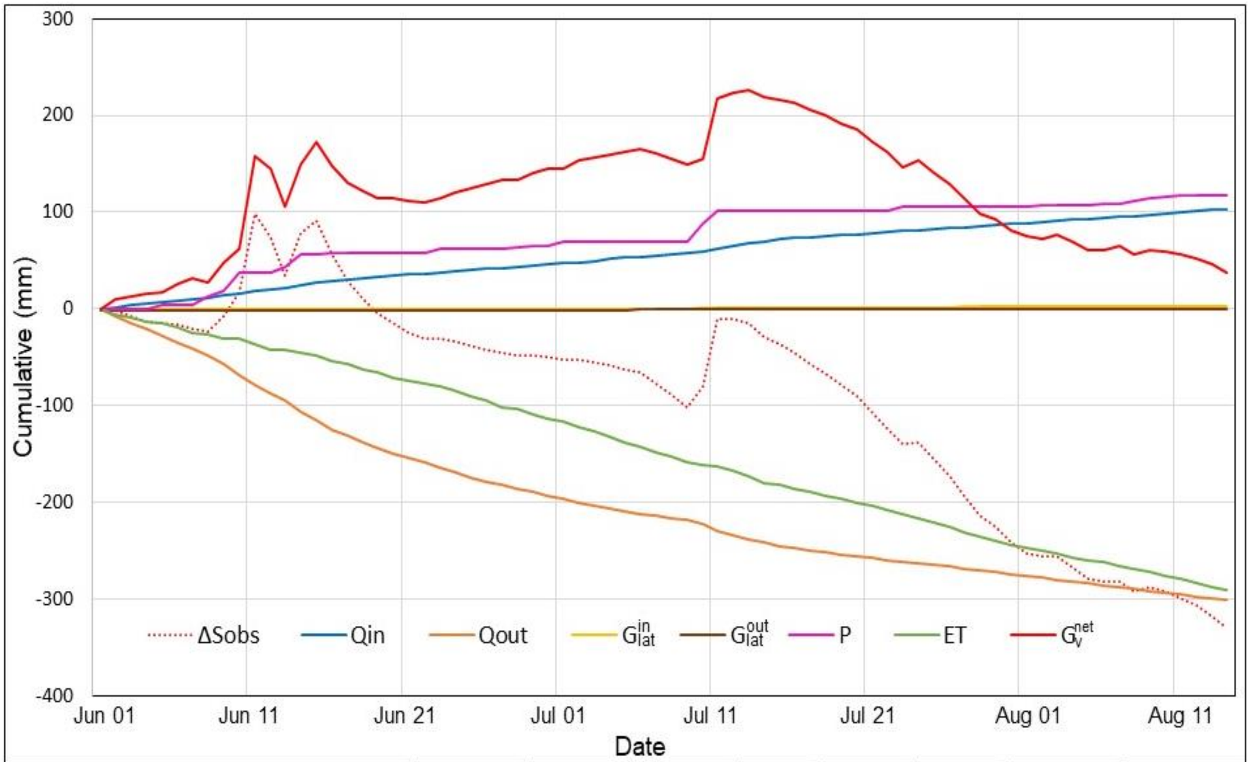


Figure 3.11 Daily change in storage (ΔS, mm/day) and the calculated net vertical groundwater exchange between the alluvial aquifer and the peat in relation to precipitation at the Sibbald Research Wetland, 1 June – 13 August 2017.

3.4.6 Vertical groundwater flux

Net vertical groundwater exchange between the peatland and the alluvial aquifer was not measured, but rather calculated as the residual of the water balance and thus includes all the error of the water balance calculations (Figure 3.11). Positive values indicate groundwater input to the peatland, whereas negative values indicate groundwater exchange to the aquifer. For most of the study period, net vertical groundwater flux ranged between 0.3-15 mm/day, averaging 3.8 mm/day. In June, vertical groundwater was primarily flowing from the alluvial aquifer towards the peat, whereas in July and August, vertical groundwater was primarily flowing from the peat to the alluvium. Additionally, three distinct periods of higher vertical groundwater flow coincided with the 8-10 June, 13-14 June and 10-11 July rainstorms. During these storms, the peatland received 130, 70, and 81.7 mm of groundwater, respectively. These large inputs of water were immediately followed by flow reversals where the peatland contributed water to the underlying alluvial aquifer. Small precipitation events in late July and August (<5 mm) also seemed to trigger small flow reversals, where flux most commonly went from the underlying alluvial aquifer to the peat aquifer. These flow reversals were largest on 12-13 June, as well as 16-20 June, when 22.9 and 61.7 mm of water moved from the peatland to the underlying aquifer. In sum, the peatland received 37.0 mm of water from the underlying alluvial aquifer over the monitoring period (Figure 3.13).



Parameter	ΔSobs	P	Q_{in}	G_{lat}^{in}	ET	Q_{out}	G_{lat}^{out}	G_v
Total Cumulative Flow (mm)	-330	118	103	3	-290	-301	0	37
Total Input (mm)	261							
Total Output (mm)	591							

Figure 3.12 Cumulative water budget of the Sibbald Research Wetland. The largest input of water to the wetland was vertical groundwater (G_v^{net}) from the exchange between the alluvial aquifer and the peatland, whereas the largest loss of water was due to evapotranspiration.

3.5 Hydrological Function

Figure 3.14 shows the daily distribution of hydrological function at Sibbald Research Wetland. From 1 June to 6 July, the peatland predominantly transmitted water to Bateman Creek Outlet (accounting for 24.4% of total flow during this period), which coincides with the period in which the peatland became frost-free. From 7 July to 13 August, the peatland predominantly contributes water to the outlet (accounting for 44.3% of flow during this period). Additionally, evapotranspiration is a critical function throughout the study period as the peatland is always losing water to the atmosphere. The ET range was 0 - 67% of daily flux. Peatland recharge was the dominant hydrological function during the storm events.

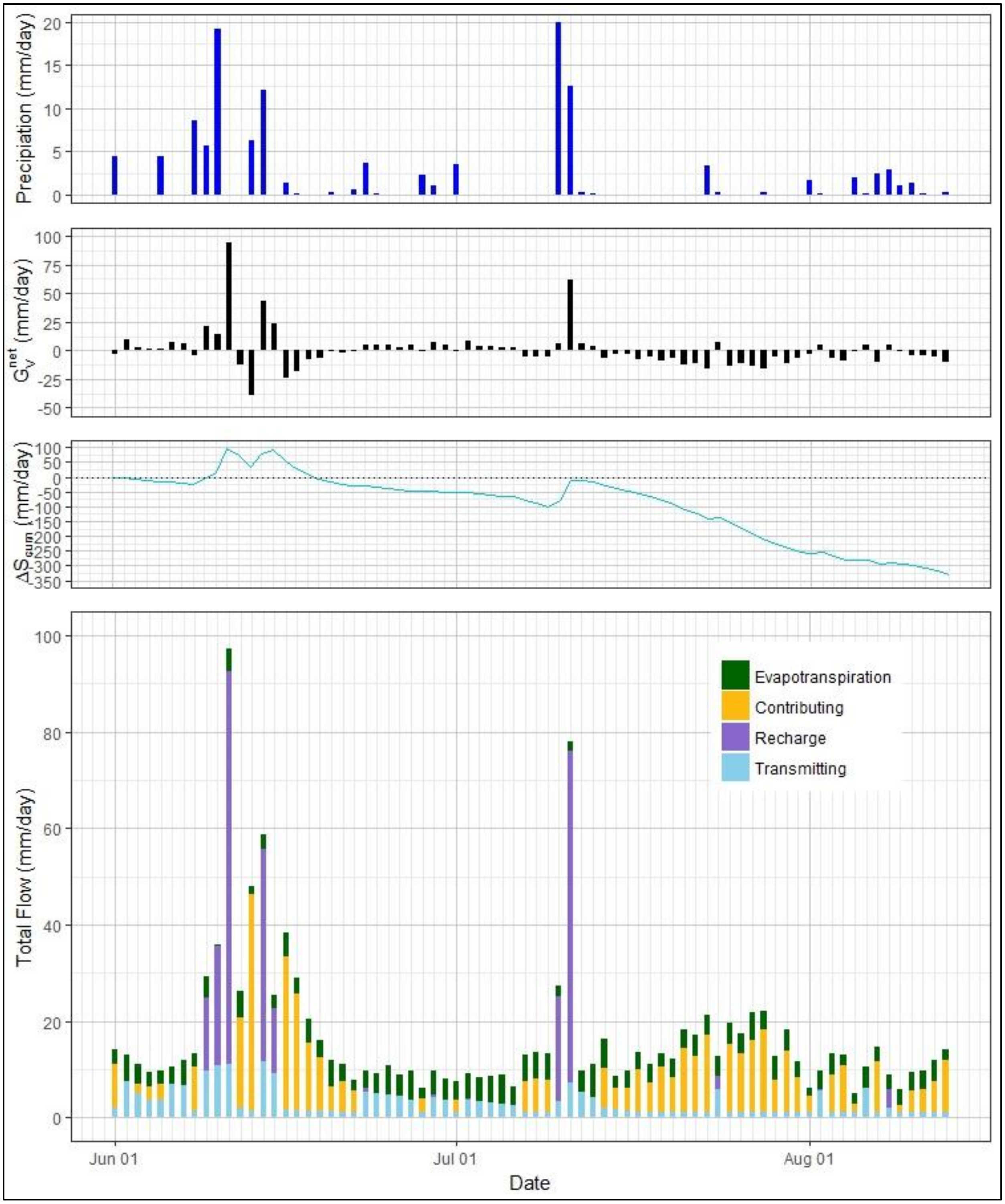


Figure 3.13 Daily distribution of wetland hydrological function at the Sibbald Research Wetland. Changes in predominant function occur during shifts in storage, precipitation and vertical groundwater fluxes. Ground frost is present in the peatland until the first week of July. After this, the peatland is frost-free.

CHAPTER 4 - DISCUSSION

Between the months of June and August 2017, the Sibbald Research Wetland regulated runoff in four different ways: it transmitted, contributed, stored and evapotranspired water. Shifts in hydrological function helped characterize the hydrological regime of the peatland and varied with changes in frost thaw, water table depth, precipitation events and vegetation productivity. Results from this study demonstrate that this mountain valley-bottom peatland can still generate runoff despite extreme drought that occurred during the monitoring period.

From 2-7 June and 23 June-6 July the dominant hydrological function of the peatland was transmission. During both periods, the same amount of runoff recorded at the outlet each day was transmitted by the peatland. The water transmitted by the peatland was mainly via streamflow as the peatland itself released little of its stored water during these two time periods. For instance, the water table at W4 only fell from 26 to 32 cm below the surface between 22 June and 6 July. The peatland released little storage during the two periods of transmission because of the presence of ground frost, which commonly occurs in northern peatlands (Glenn and Woo, 1997; Quinton et al., 2003). Frost was widespread during the entire month of June. Frost depth ranged between 17-50 cm below the surface, with frozen areas closest to the surface occurring at the beginning of June, and frost gradually receding towards the end of the month. The frost impeded internal stores of water from exiting the peatland, by reducing saturated hydraulic conductivity (Hogan et. al, 2006). For instance, from 1-7 June, daily changes in storage averaged only -3.3 mm/day for 2-7 June and -2.5 mm/day for 23 June- 6 July. Under these frozen conditions, discharge at Bateman Creek Outlet decreased.

From 14 July until the end of the study, the peatland's dominant hydrological function was contribution. As the peatland was mainly frost-free during this time, the system was able to generate runoff by draining internal stores of water, which augmented streamflow (Wright et al., 2009). Water table levels declined during this time, as did total storage. From 11 July to 13 August, the water table dropped 0.74 m at W4 in the north end, and 0.36 m at W61 and W62 in the south end. The rate of storage loss was nearly four times higher than when the peatland was frozen, averaging -11.1 mm/day. The release of stored water from the peatland was critical in maintaining streamflow in Bateman Creek during the dry period. Others, for example Glenn and Woo (1997), have also shown that runoff generation from northern peatlands becomes increasingly controlled by intra-wetland processes under low flow conditions.

Ground thaw was not uniform throughout the peatland, so frost table depth became progressively heterogenous with time. For instance, 7 June frost table surveys indicate that frost was still present in the upper 50 cm of peat throughout the peatland. Frost was mainly absent on 28 June but in the northern part of the wetland until 28 July. Heterogenous frost thickness in peat during melt has been documented in other studies (e.g. Guan et al., 2010). Furthermore, frost tables increase the hydrostatic pressure of the underlying groundwater. As a result, groundwater can seep out of the peatland in areas where frost is absent (Carey and Woo, 2001). Variations in frost table depth results in fluctuations in available storage space (Woo and Xia, 1995) which likely influenced outflow to Bateman Creek Outlet, especially during precipitation events.

Ground frost, which did not fully melt until July 7, was also important in regulating stormflow response. During the study period, there were three sizable rainstorms. The peatland experienced a hydrological function switch from peatland recharge early in each rainfall event contribution later in each event. However, the two storms (8-10 June and 13-14 June) that occurred when the ground was frozen incited different stormflow responses than the storm that occurred when the ground was thawed (10-11 July). The 8-10 June rainstorm had nearly double the amount of precipitation than the 13-14 June rainstorm, yet Bateman Creek discharge during the second storm was much higher. Furthermore, the 10-11 July event, which had nearly the same amount of precipitation as the 8-10 June event yielded only half the amount of discharge. The relationship between precipitation and outlet discharge can be explained by the frost table dynamics and antecedent moisture conditions of the peatland.

During the first half of June, when frost tables were high, the first large rainstorm easily and quickly satisfied storage. High antecedent wetness conditions also factored into the larger 13-14 June stormflow response. Accordingly, runoff ratios for this storm indicate that much more precipitation passed through Bateman Creek Outlet than precipitation during the 8-10 June storm. High water tables during this time produced a week-long stormflow response, when the peatland contributed water to Bateman Creek Outlet. This process has been seen in several northern wetlands (e.g. Woo and Steer, 1983; Hayashi et al., 2007) including a fen in the continuous permafrost zone of the eastern Northwest Territories (Roulet and Woo, 1986). There, the fen was found to be a poor regulator of flow during spring precipitation and snowmelt when permafrost, found 55 cm below the peat surface, created an impermeable barrier to deeper groundwater. The fen was quickly flooded during spring inflows as frost presence created limited storage. Excess water began to leave the peatland by flowing overtop of the saturated subsurface, further decreasing storage. In contrast, the large 10-11 July rainstorm at the Sibbald Research Wetland produced significantly less runoff than either of the two June storms even though precipitation amount was nearly the same as the 8-10 June event. During the storm the peatland was frost-free, and absence of frost increased the amount of available peatland storage, allowing more precipitation to percolate into the peat rather than running off into the outlet

That ground frost plays a critical role in regulating peatland hydrological function is not a new concept. Studies across northern landscapes have shown that frost limits peatland storage capacity and the ability for the peatland regulate streamflow (Roulet and Woo, 1986; Glenn and Woo, 1997; Quinton et al., 2003; Wright et al., 2009). In mountain environments, however, peatland runoff regulation processes may be more complicated as frost is not always present in peatlands of different elevation. This phenomenon may differ from peatlands in other ecozones where soils are either permanently or seasonally frozen across large spatial scales. Indeed, peatlands in the Canadian Rocky Mountains occur across a large elevation gradient (Morrison et. al.,2015) where deeper snowpacks (>30 cm) and colder temperatures are found at high elevations (Hayashi et al., 2003; Millar et al., 2017). Deep snowpacks have been shown to insolate peatlands from frigid conditions, preventing the formation of frost tables (Fuss et al., 2016). Conversely, thin snowpacks (<30 cm) have been linked to deeper frost tables, as the insulative barrier between the ground and freezing air temperatures is thinner (Hayashi et al., 2003; Hayashi, 2013; Fuss et al., 2016). The effect of thin snowpacks can be seen at the Sibbald Research Wetland. Here, average monthly snowpack thickness from November-April 2017 was ≤30 cm and frost table surveys provide evidence of a well-established, deep frost table, in turn affecting runoff regulation in the peatland. A nearby peatland in the alpine though has soils that remain thawed through winter (Mercer, 2018). Future research might therefore entail understanding the impact of frozen ground on hydrological function in peatlands of different elevation. Additionally, as air temperatures gradually start to increase under warming climates, understanding the effect of variable snowpack thickness across a range of elevations would aid in predicting future runoff generation patterns in mountain peatlands.

Little lateral groundwater flow exchange between the hillslopes and the peatland was captured by the water budget. This is a bit of an unusual observation. The peatland (and its underlying alluvial aquifer) are sloping, where the north end of the peatland is 32m higher than the south end. Sloping mountain peatlands usually show storing lateral hydraulic gradients (Woods et al., 2006; Millar et al., 2018). However, topography in mountain regions has been shown to influence the shape of the water table, which takes on a similar shape to the underlying landscape unit, thus influencing groundwater exchange between landscape units (Forester and Smith, 1988; Gleeson and Manning, 2008). A recent three-dimensional simulation of groundwater flow in mountainous regions further confirmed topographically driven recharge zones (Welch and Allen, 2012). Results from the Welch and Allen simulation suggest that groundwater recharge and discharge in valley-bottoms reflects the topography of the valley slopes, with baseflow being produced at valley outlets. Thus, transverse groundwater exchange between the underlying alluvial aquifer and the peat likely attributed to the transmission and contribution of water to Bateman Creek Outlet. Given the lack of lateral groundwater movement recorded in the water budget, transverse flow is likely to have been captured in the vertical groundwater calculation. As the latter was calculated as the remainder of the water budget, the magnitude of G_v^{net}depends on the magnitude of AET. G_v^{net} proved to be an important source of stored water to the peatland. It was the largest source input after precipitation and incoming streamflow.

A shortcoming of the Spence et al. (2001) hydrological function model is that evapotranspiration is embedded in the storage function, meaning the model assumes that peatland storage is only lost through runoff generation to a stream. However, evapotranspiration is an important part of the wetland water balance, and has been well documented across the Prairie Pothole, Boreal and Arctic regions as the primary pathway of water loss from wetlands in the spring and summer (Carey and Woo, 2001; van der Kamp et al., 2003; Wright et al., 2009; Phillips et al., 2011; Tardif et al., 2015). To better account for water loss from the peatland, an additional hydrological function was added to the model: evapotranspiration. The addition of evapotranspiration to the hydrological function model permits a more realistic understanding of the water release functions – water loss to the stream (via transmission or contribution) or water loss to the atmosphere. With the addition of evapotranspiration to the hydrological functions model, actual storage is thus differentiated from evaporative loss.

Evapotranspiration proved to be the second largest water loss (290 mm of water for the entire study period) at the Sibbald Research Wetland after outgoing streamflow. Daily evapotranspiration was reasonably constant throughout the study period. Further, the water table diurnally fluctuated, suggesting plant reliance on peat groundwater (Lautz, 2008). That evapotranspiration was affected by ground frost has been suggested in other northern peatlands (Petrone et al., 2008), although further research on the dynamics between frost and ET is required. Despite the importance of ET to the overall water budget for the peatland, only during the 23 June – 6 July period did evapotranspiration become the predominant function of the system. Interestingly, this period occurred when frost was still widespread in the peatland. ET tapered off late into the study as water tables fell below 0.91 m.

Not captured in the study was contributed water from beaver ponds during the storm events. It is possible that during the two large precipitation events, storage in the beaver ponds was satisfied, creating overflow that spilled beside the dams during peak flow. Water from the beaver ponds may have bypassed the dams by way of throughflow or escaping through weakened areas of dam construction (Woo and Waddington, 1990; Westbrook et al., 2006). The influence of beaver dam overflow on streamflow during events has previously been described. Woo and Waddington (1990) recorded the discharge of an Arctic stream regulated by beaver dams after a major precipitation event. Once pond storage is exceeded, water overflows the dams. Woo and Waddington (1990) noted that it took five days for downstream river discharge to reach pre-event levels, which was slower than streams lacking beaver ponds. The Sibbald peatland has many beaver ponds (Morrison et al., 2015; Karran et al., 2018) which likely played a role in the observed gradual return to pre-event discharge.

CHAPTER 5 - CONCLUSIONS

5.1 Summary of Findings

To examine the role of large, valley-bottom, mountain peatlands in regulating streamflow, a water balance approach along with a modified version of the Spence (2007) wetland hydrological functions model were used. This model states that wetlands can store, transmit, and contribute water at any one time. From 1 June to 13 August, during a regional period of drought, daily predominant hydrological function at the Sibbald Research Wetland was characterized and controls on functional changes were explored. Results show that there were two main functions throughout the study period. For the month of June until the first week of July, the peatland predominantly transmitted water to its outlet. Throughout this time, the peatland released very little water from its internal stores. Thereafter, the peatland predominantly contributed water to Bateman Creek Outlet by releasing its internal stores of water. Temporary switches in hydrological function occurred during large rainstorms, wherein the peatland first briefly recharged then contributed to streamflow.

The frost table was the most important initiator of functional change. During the month of June, frost tables were within the upper 50 cm of peat, impeding stored water from being released into Bateman Creek. During this time, the peatland poorly regulated flow as any incoming water was not able to percolate into the deeper peat in appreciable quantities and so did not become part of the deeper ground storage. Incoming water was transmitted by the peatland or rapidly contributed to the stream in the form of shallow subsurface or overland flow. In the absence of frost, the peatland predominantly contributed water to Bateman Creek. From the second week of July to the end of the study, the peatland released water from its internal stores, continuously supplying Bateman Creek with runoff despite extreme drought. Indeed, during a summer where many surrounding streams ran dry, Bateman Creek was able to maintain baseflow.

A shortcoming of the Spence (2007) model is that peatland storage can only be lost through runoff generation. I modified the model to also have evapotranspiration (ET) as a hydrological function. ET has repeatedly been shown in the literature to be an important pathway for water loss from peatlands (Carey and Woo, 2001; Quinton et al., 2003; Waddington et al., 2009). At the Sibbald Research peatland, ET was often the predominant driver of water loss in the system, especially during periods of high frost table. This hydrological pathway had a regulating effect on how much water was contributed to the peatland outlet. To add an ET hydrological function to the Spence (2007) model, I delineated the storage function into peatland recharge and ET. The recharge function occurred when daily changes in storage were greater than zero, and ET occurred when changes in storage decreased (lower than zero). The ET function was rarely the dominant function during the study period. However, it was consistently important in decreasing daily peatland storage.

5.2 Limitations of study

This study has a few limitations, mostly owing to the timing of the research, as well as the large size of the peatland. As the study period was limited to three months, peatland function under a range of climate conditions was not captured. In fact, the study period presented only severe drought conditions. Although this research was able to capture the runoff regime of the peatland in severe conditions, the results from this study do not show the behaviour of the peatland in average or wet antecedent conditions. However, the study was able to provide insight into the role of peatlands in maintaining low flows despite its shortness, which is a topic of considerable interest (Smakhtin, 2001). Secondly, the calculation of observed storage assumed an average specific yield for the entire peat column. This number usually declines with depth of peat (Kettridge et al., 2015). However, I used a single value, which was only representative of the shallow peat found at the Sibbald Research Wetland. As less water is needed in the lower peat column to incite changes in water table depth, using a single specific yield value designed for the upper peat column adds uncertainty to the calculation of storage. Thirdly, open water storage (mainly beaver ponds) accounted for 12% of the total area of the peatland. How beaver ponds influence peatland hydrological function was not explicitly accounted for in this study as their water levels were not measured. However, they are an important part of water storage within the peatland and so should be studied to see how they influence hydrological function. Fourthly, relocation of the meteorological station midway through the study period was problematic for volumetric water content. Soil moisture in the new location was higher than that of the old location, and data were not collected at both locations simultaneously for use in data correction. This creates uncertainty in the volumetric water content data, but fortunately, has little impact on total water storage. Finally, a large degree of uncertainly was captured in the ET and G_v^{net} calculations. As eddy covariance was not available for this study, the Penman-Monteith (PM) method was used to calculate actual ET. While Penman-Monteith is a preferred approach for calculating actual ET, some parameters used in the calculation are uncertain. Specifically, leaf area index for sedge and willow were not measured on-site, and were instead taken from other studies, where environmental conditions were similar to but not the same as those of the Sibbald Research Wetland. With the installation of an eddy covariance tower at this site in the summer of 2018, uncertainty of future estimations of evapotranspiration should be reduced.

5.3 Research implications and future work

My research on hydrological function at the Sibbald Research Wetland resulted in two major findings. Firstly, the peatland constantly supplied Bateman Creek with runoff despite drought. Notwithstanding the dry conditions of the region, and the fact that streams flowing into the peatland ran dry halfway through the summer, Bateman Creek continued to flow throughout the monitoring period. Streamflow was supplied by the internally generated water of the peatland, supplemented by groundwater inputs from the underlying alluvial aquifer. This resulted in a water table decline of close to 1m. Secondly, peatland hydrological function was largely controlled by the presence of ground frost. In periods where frost was within the upper 50 cm of peat, the peatland served as an intermediary landscape unit, directly transmitting inflow to Bateman Creek Outlet. Very little runoff came from storage, as frost impeded access to the peatland's internal supplies of water. However, in times of absent frost, the peatland contributed stored water to the outlet, further depleting the water table. This research therefore suggests that these large mountain reservoirs are able to maintain baseflows even when other types of reservoirs (such as the snowpack) are depleted. Additionally, mountain valley-bottom peatland hydrology is controlled by seasonality, where frost thaw can largely regulate the release of water.

Although results from this study provide insight into the runoff regime of a mountain-valley bottom peatland and its controls, questions remain on the applicability of this research under different climate conditions. Does the runoff regime of the Sibbald Research Wetland follow the same pattern of hydrological function under atmospheric conditions similar to those of the 10-year climate norm? Furthermore, how does a peatland's runoff regime differ in time of extreme wetness, such as the 2013 flood? Future research should involve capturing peatland hydrological function under a range of climate conditions.

Mountain regions are known for their extreme and varied climates owing to their complex topography. In general, precipitation increases with elevation, while air temperatures decrease with elevation (Baron and Denning, 1993). Precipitation may also differ between sites depending on peatland location (Evans et al., 1999; Chimner et al., 2010). Areas within the rain shadow of a mountain, for example, will receive less rain than areas outside of the rain shadow. Hence, peatlands may exhibit varying hydrological function based on their topographical context and location. For instance, peatlands at low elevations are known to have deep frost tables in winter and spring due to thin snowpacks, whereas peatlands of higher elevation are known to be insulated by deep snowpacks, providing little chance for seasonal ground frost to form (Millar et al., 2017). Runoff regimes in mountain peatlands of different elevation are likely to be controlled by site-specific conditions such depending on the topography of the area (Millar et al., 2018). Future research should examine the hydrological function of peatlands at different elevations and different precipitation regimes to understand the characteristics and controls on regimes in areas of different elevations including the insolating properties of snow on the formation of the frost table.

Finally, multiple beaver dams and ponds were present before this research (Janzen and Westbrook, 2011) and several have appeared at the Sibbald Research Wetland since the conclusion of fieldwork for this study. It was an anomaly that there were few beaver during my study period. Beaver are known to influence the water table of peatlands. For instance, water tables at the Sibbald Research Wetland were found to elevate and stabilize water tables within a 150 m radius of a beaver dam (Karran et. al., 2018). Beaver dams also regulate water table decline and baseflows at peatland outlets (Woo and Waddington, 1990; Westbrook et al., 2006). Thus, the runoff regime at the Sibbald Research Wetland has possibly been altered since my study. Future research might involve understanding the effect of well-established beaver ponds on the hydrological function of the Sibbald Research Wetland as well as other peatlands.

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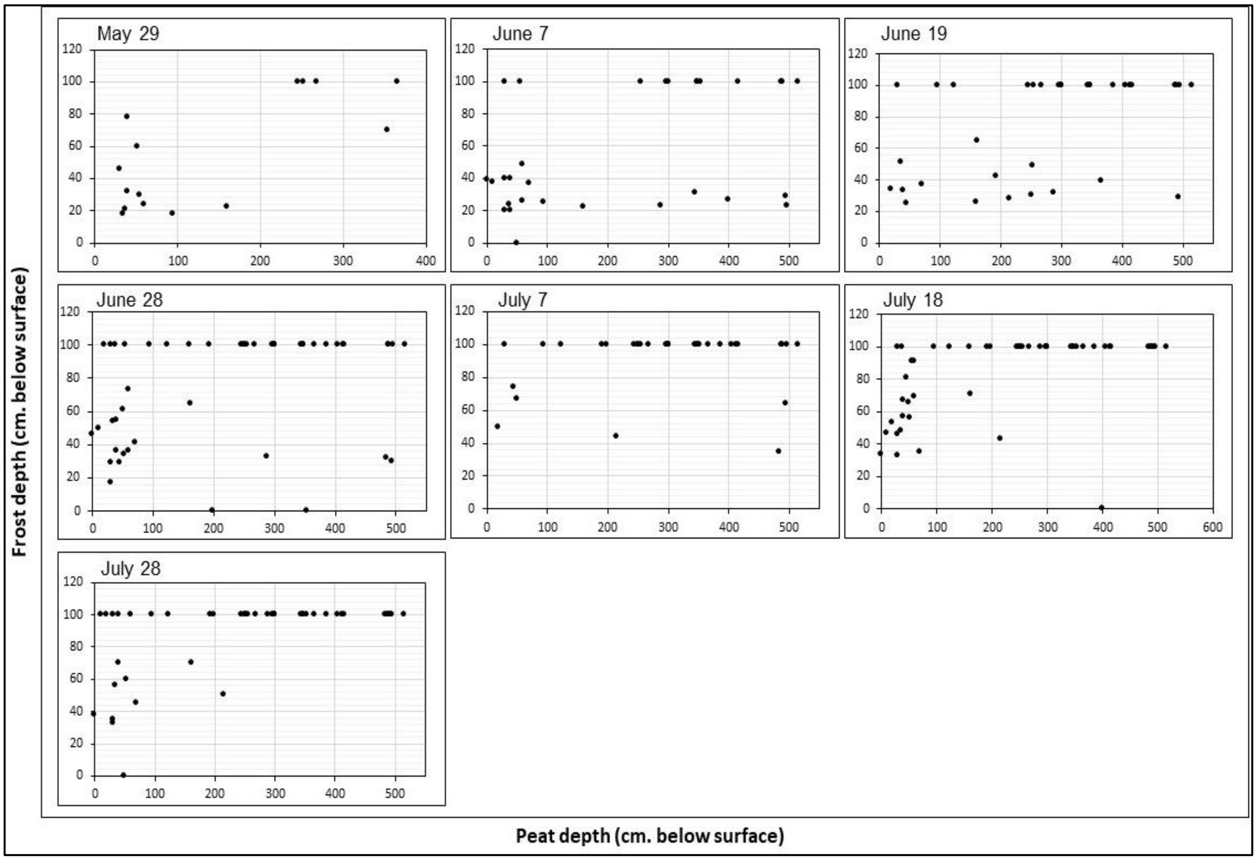

Figure A.1 Plots of peat depth (cm below the surface) vs. frost depth (cm below the surface) at the Sibbald Research Wetland. In general, frost table depth and peat depth are proportional: frost tables and shallowest in areas where peat in shallowest, and vice versa.

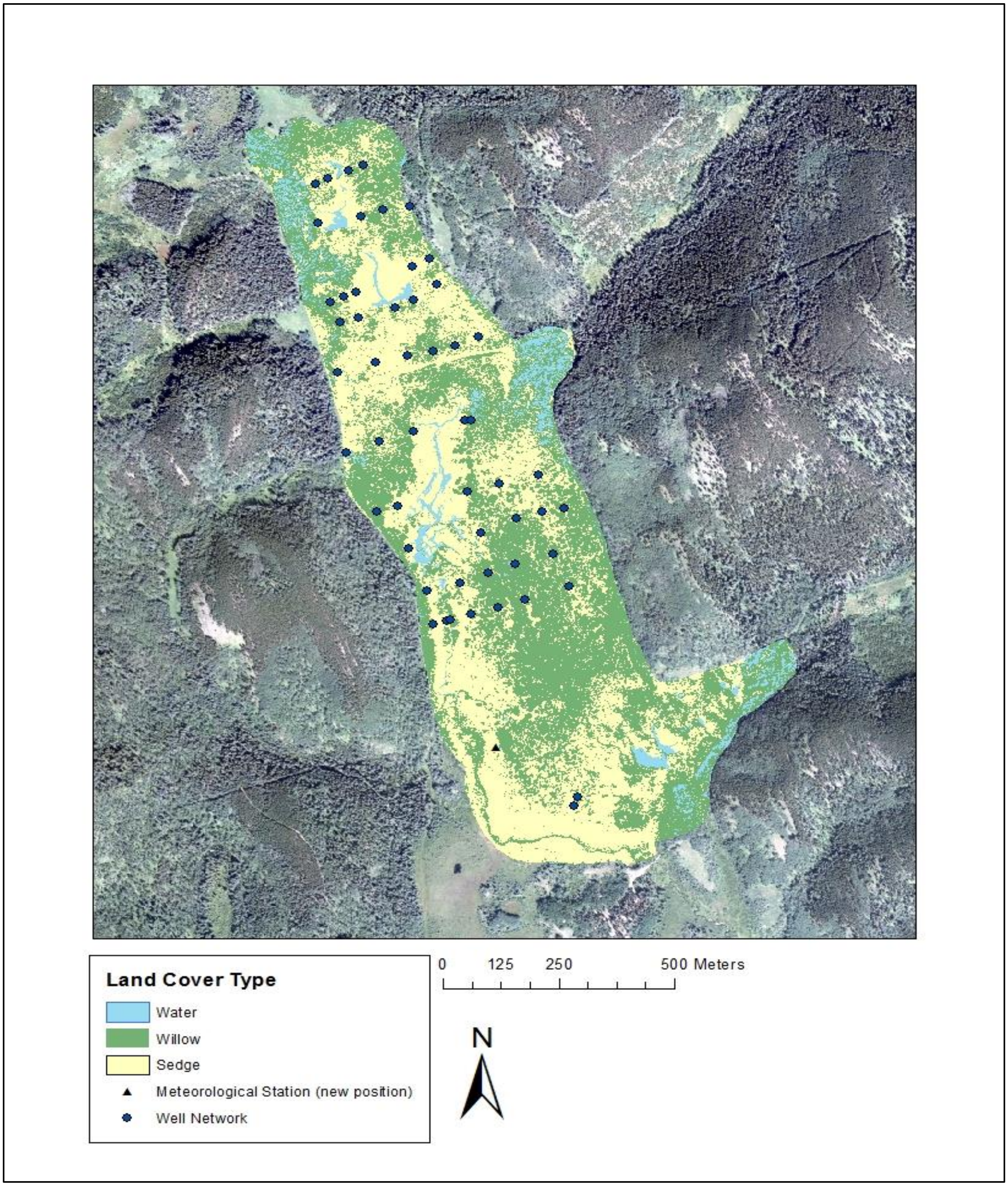

Figure B.1 Maximum likelihood classification of the Sibbald Research Wetland with 2013 imagery. Here, land-cover is divided into three main types: sedge, willow and open water. It was estimated that sedge, willow and open water made up 65%, 23% and 12%.

Table B.1 List of parameters used in calculating evapotranspiration with the Penman-Monteith approach

Parameter	Value
Maximum aerodynamic roughness height	2.5 m
Height of wind measurement	3.05 m
Height of vapor density measurement	2.36 m
Mean height of sedge canopy	0.5 m
Displacement height of sedge	0.3 m
Roughness length of momentum (sedge)	0.0625 m
Roughness length of water vapor (sedge)	0.00625 m
Mean height of willow	3 m
Roughness length of momentum (willow)	0.375 m
Roughness length of water vapor (willow)	0.0375
LAI sedge	0.62
LAI willow	2.3

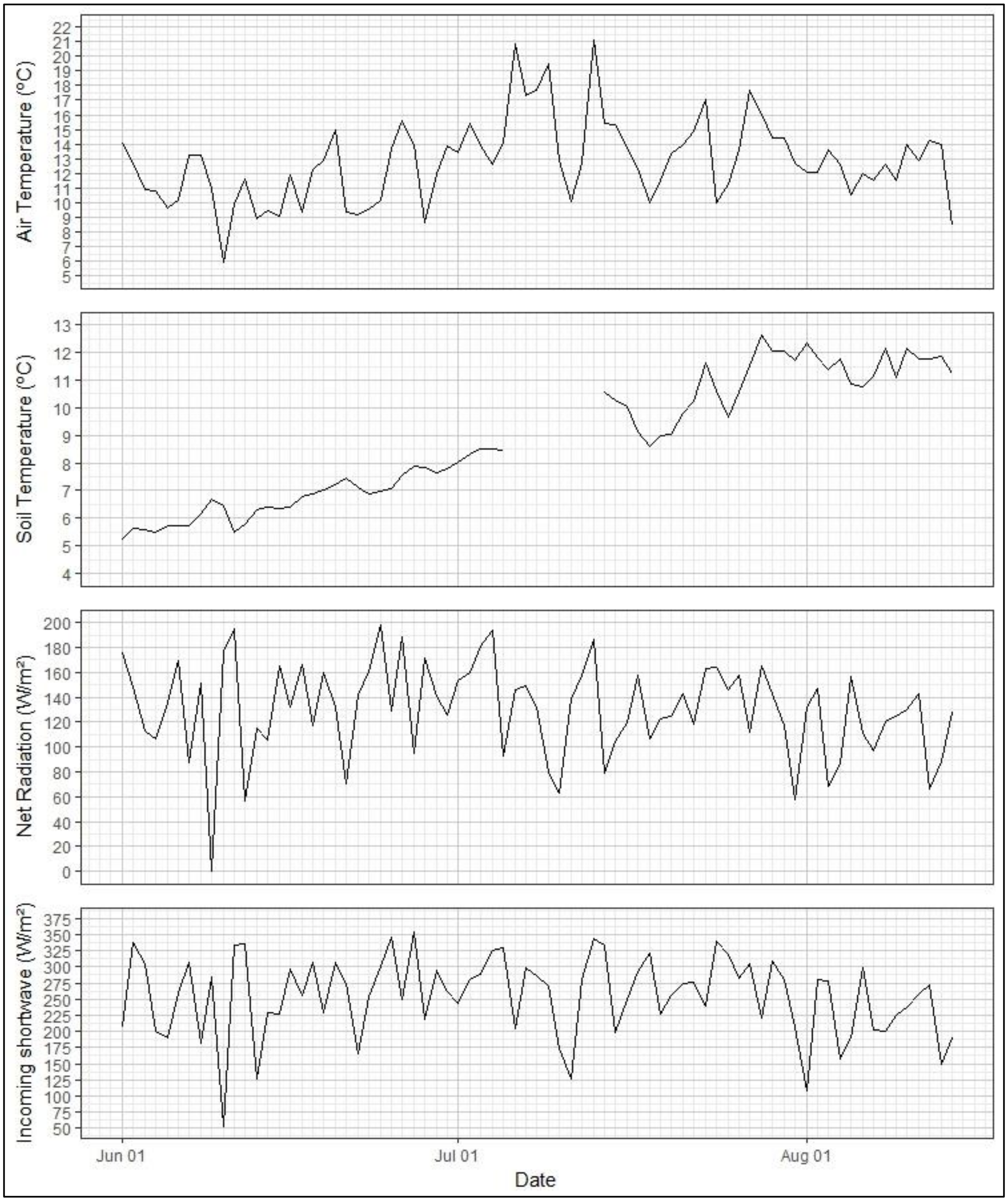

Figure C.1 Daily values of incoming solar radiation (W/m²), net radiation (W/m²), soil temperature (ºC) and air temperature (ºC) at the Sibbald Research Wetland from 1 June to 13 August, 2017. The gap in the soil temperature record indicated the period when the meteorological station was moved.

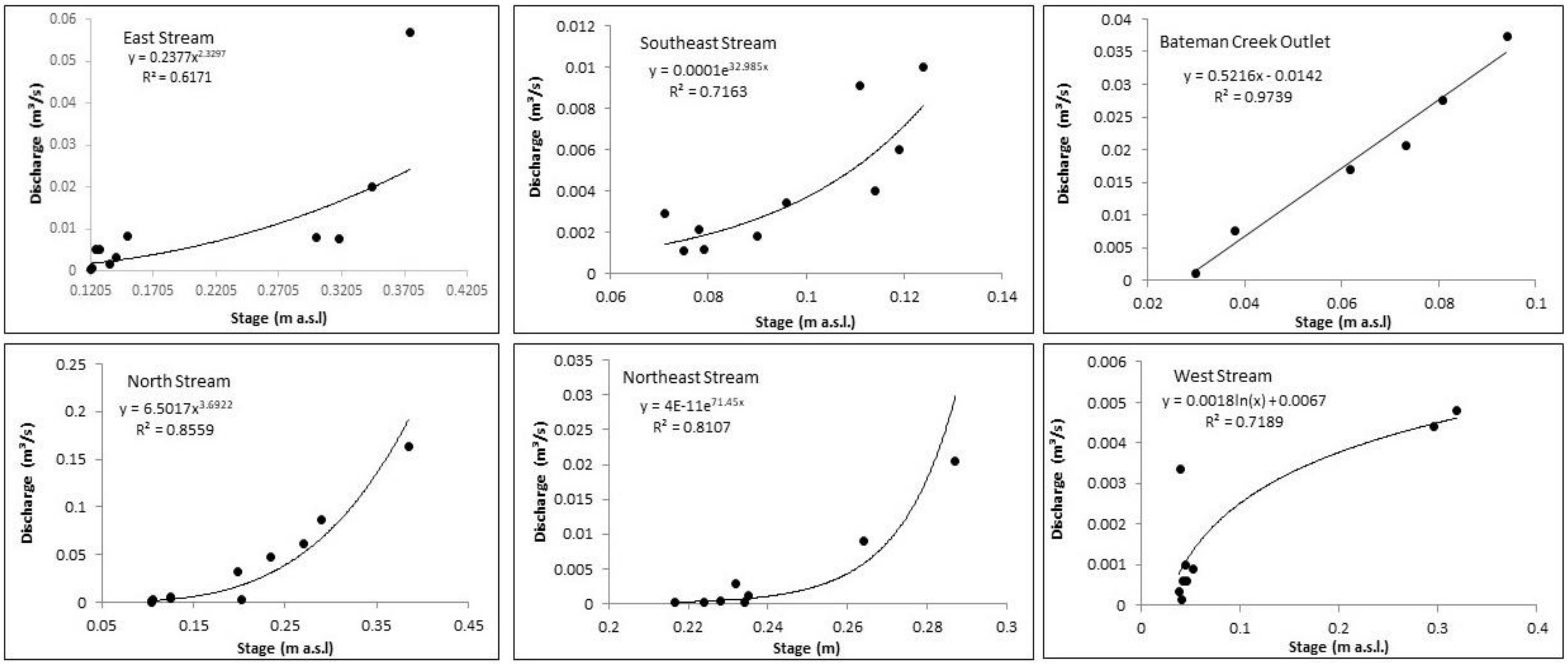

Figure D.1 Rating curves with respective equations for the east, southeast, north and northeast streams, as well as Bateman Creek Outlet at the Sibbald Research Wetland. Both the area of the peatland and watershed for the outlet stream gauge are 0.71 km² and 9.3 km², respectively.

APPENDIX E - METADATA

Table E.1 Peatland measurements

Parameter	Value
Area of peatland (km^2)	0.71
Area of watershed (km^2)	9.3
Peat depth at meteorological station (new location, cm)	80
Peat depth at meteorological station (old location, cm)	65
Average peat depth of peatland (cm)	167.5
Average peat thickness at south end (cm)	80
Average peat thickness at north end (cm)	78
Width of inlet (m)	214.9
Width of outlet (m)	16.9

Table E.2 Saturated hydraulic conductivities taken at wells 4, 7, 61 and 62 using the Hvorselv method at the Sibbald Research Wetland.

Well	Hydraulic conductivity (m/s)
W4	7.16×10^{-7}
W7	2.88×10^{-6}
W61	6.83×10^{-6}
W62	1.54×10^{-6}

Table E.3 rtkGPS coordinates of all monitored groundwater wells during the 2017 study at the Sibbald Research Wetland.

	GPS coordinates				
Well #	**Latitude**	**Longitude**	**UTM X (Easting)**	**UTM Y (Northing)**	**Elevation (m a.s.l)**
4	51° 04' 05.41928" N	114° 52' 20.93354" W	649061.1	5659559.2	1485.0711
7	51° 04' 02.01757" N	114° 52' 18.66188" W	649108.3	5659455.4	1482.6802
61	51° 03' 16.78534" N	114° 51' 57.11016" W	649568.2	5658070.5	1470.2545
62	51° 03' 16.10656" N	114° 51' 57.52933" W	649560.7	5658049.3	1470.0066
Ppeat	51° 03' 53.35791" N	114° 52' 20.49436" W	649080.4	5659186.9	1479.3744
PAlluvium	51° 03' 53.36007" N	114° 52' 20.49706" W	649080.3	5659187	1479.456
Met station	51° 03' 20.70035" N	114° 52' 06.23397" W	649387.1	5658186.3	1469.865

Table E.4 Pipe configurations of all monitored groundwater wells at the Sibbald Research Wetland during the 2017 study.

	Pipe configurations						
Well #	**Length of pipe (cm)**	**Length of screen (cm)**	**Distance from screen to top of pipe (cm)**	**Length of pipe above ground (cm)**	**Inner diameter (cm)**	**Outter diameter (cm)**	**Diameter of auger hole (cm)**
4	150.8	110	35.8	31.8	4.5	4.8	5
7	184	110	67	72.1	4.5	4.8	5
61	143.9	132.5	7.1	24.6	2.5	2.8	5
62	160.6	157.2	3.2	33.9	2.5	2.8	5
Ppeat	155.6	15	134.6	59.4	4.5	4.8	5
PAlluvium	243.84	81.1	152.4	63	5.5	5.8	5.8

Table E.5 Meteorological station metadata at the Sibbald Research Wetland for the 2017 study period

Manufacturer	Variable	Variable Name	Units	Measurement Type	Height Above Ground (m)	Serial Number
Rotronic	Air Temperature	AirTemp	º C	Average	2.36	137154
	RH	RH	%	Sample		
RM Young	Wind Speed	WindSpeed	m/s	Windvector	3.05	80589
	Wind Direction	WindDir	degrees	Windvector		
Campbell Scientific	Soil Temperature	SoilTemp	º C	Average		C7078
Texas Electronic	Rainfall	Precip	mm	Total	2.95	42886-607
Kipp & Zonen	Net Radiation	NR_Wm2	W/m²	Average	1.57	N/A
Campbell Scientific Canada	Snow Depth	SnowDepth	m	Average	1.17	C1869
	Signal Quality	SignalQuality	unitless	Average		
Radiation Energy Balance Systems	Soil Heat Flux	HFP	W/m²	Average		Job# 159442
Omega	Soil Temperature	SoilTemp_Sur	º C	Average		N/A
	Soil Temperature	SoilTemp_Deep	º C	Average		
Apogee	Incoming Shortwave Radiation	K_In	W/m²	Average		15413
Campbell Scientific	Volumetric Water Content	VWC_shallow	%	Sample	-0.25	300457
	Period	Period_shallow	µs	Sample		300457
	Volumetric Water Content	VWC_deep	%	Sample	-0.50	Job# 159442
	Period	Period_deep	µs	Sample		Job# 159442
Campbell Scientific	Battery Voltage	Batt_Volt	volts	Minimum		12430
	Battery Voltage	Batt_Volt	volts	Maximum		
	Panel Temperature	PTemp	ºC	Minimum		
	Panel Temperature	PTemp	ºC	Maximum		

APPENDIX F - *ANTECEDENT PRECIPITATION INDEX*

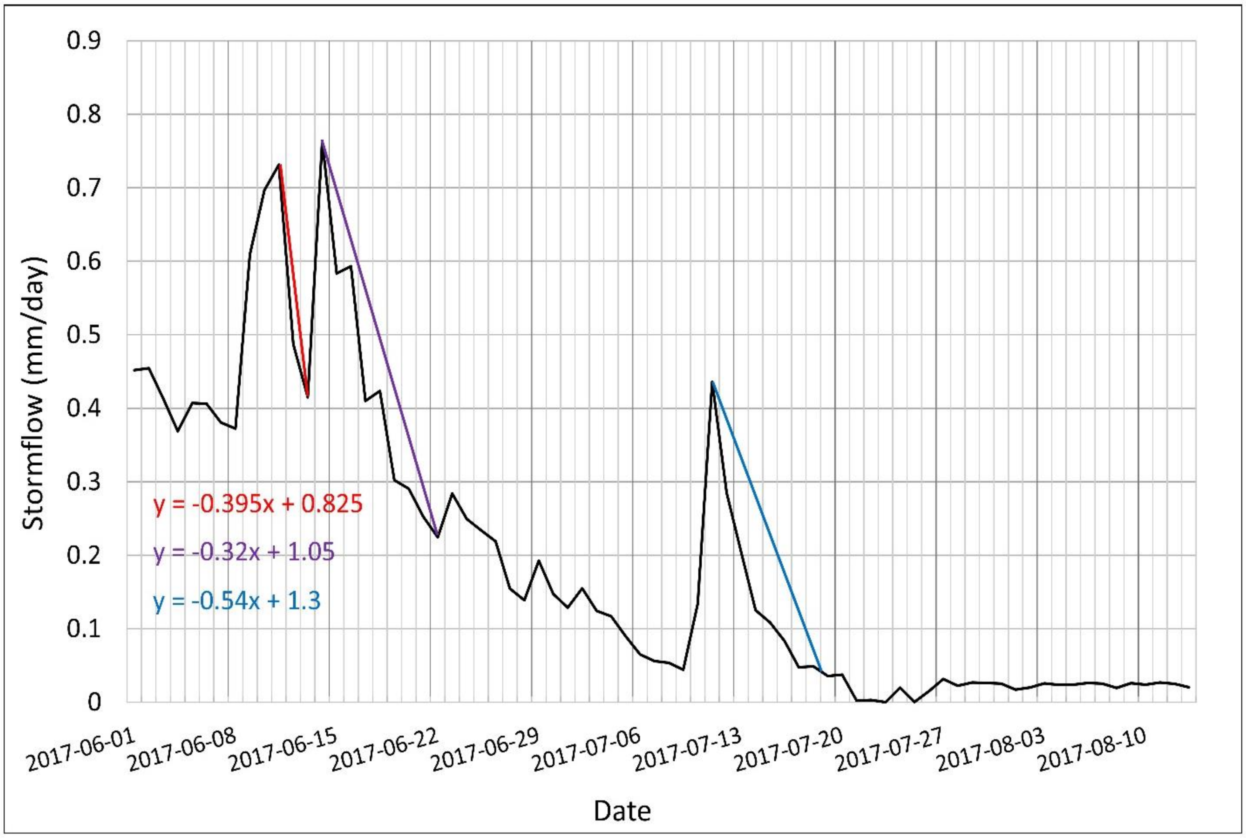

Figure F.1 Stormflow recession limbs at Bateman Creek Outlet during the three largest rain events at the Sibbald Research Wetland. The decay factor used in calculating the antecedent precipitation index decay factor was taken as the average of all three regression slopes (0.4).